Chapter 6
Textural and chemical characteristics of
diagenetic and hydrothermal alteration in glassy
volcanic rocks

6.1 Introduction

Although the characteristics of hydrothermal alteration associated with VHMS deposits have been
studied in detail (Iijima, 1974; Ishikawa et al., 1976; Eastoe et al., 1987; Gemmell and Large, 1992;
Large, 1992; Offler and Whitford, 1992; Barrett and MacLean, 1994), much less is known about the
effects of regional diagenetic alteration on glassy volcanic rocks. Diagenetic alteration can significantly
alter the textures and compositions of glassy volcanic rocks. Distinguishing between the products of
diagenetic and hydrothermal alteration is essential for efficient exploration but can be difficult. This
chapter develops textural and compositional criteria that help discriminate between diagenetic and
hydrothermal alteration facies in the northern Central Volcanic Complex. These criteria could help in
the discrimination of prospective from unprospective areas for VHMS mineralisation in other submarine
volcanic successions.

The northern Central Volcanic Complex is dominated by originally glassy volcanic rocks that
now have textures and compositions that reflect subsequent modification by diagenesis, hydrothermal
alteration, regional metamorphism and deformation. This chapter focuses on regional alteration in
the northern Central Volcanic Complex east of the Mount Black Fault. Here the northern Central
Volcanic Complex includes the Sterling Valley Volcanics, the Mount Black Formation (formerly part
of the Mount Black Volcanics) and the Kershaw Pumice Formation (formerly part of the Mount
Black Volcanics) (Fig. 4.2). Alteration facies are described in terms of the alteration mineral assemblage,
distribution, texture, intensity, overprinting relationships and geochemistry. The influence of primary
volcanic facies characteristics, especially porosity, permeability and the presence of volcanic glass, on
the distribution and texture of the alteration facies is considered. Diagenetic alteration in the northern
Central Volcanic Complex is compared to footwall hydrothermal alteration associated with the Rosebery
VHMS deposit. Specific criteria that discriminate the products of diagenetic from hydrothermal
alteration have been established.

6.2 Alteration facies in the northern Central Volcanic Complex

The present mineral assemblages and textures in the northern Central Volcanic Complex were produced
by a series of processes, including regional diagenetic alteration, compaction, metamorphism,
deformation and local hydrothermal alteration. The term “alteration facies” is used to describe a rock
unit with a distinctive mineralogy and texture. Different alteration facies have been defined on the
basis of mineralogy, texture, overprinting relationships and composition. The principal alteration
facies assemblages are (Table 6.1, Fig. 6.1): sericite, carbonate, feldspar-quartz-sericite, chlorite-sericite,
Table 6.1: Characteristics of the main alteration facies in the northern Central Volcanic Complex. 
musc = muscovite, cc = calcite, qtz = quartz, fld = feldspar, chl = chlorite, py = pyrite, 
ser = sericite, hem = hematite, mag = magnetite, ep = epidote, hbl = hornblende, px = pyroxene.

<table>
<thead>
<tr>
<th>Alteration facies</th>
<th>Assemblage</th>
<th>Texture</th>
<th>Timing</th>
<th>Mass changes</th>
<th>Total mass change</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>sericite</td>
<td>musc</td>
<td>coated glass surfaces, perlitic fractures, vesicles, sphelesites, phenocrysts and fractures</td>
<td>pre–S₁</td>
<td>gain K₂O, Al₂O₃, CaO, Rb, Sr, loss Na₂O, SiO₂, Sr</td>
<td>-9 %</td>
<td>hydration or diagenetic</td>
</tr>
<tr>
<td>carbonate feldspar-quartz-sericite</td>
<td>cc? qz, Na- or K-fld, musc</td>
<td>coated glass surfaces overgrowth on plagioclase, replaced plagioclase and glass, filled fractures and vesicles</td>
<td>pre–S₁ post–ser</td>
<td>gain CaO, total Ca²⁺, respective Na₂O/K₂O exchange, gain SiO₂, CaO, Al₂O₃, Sr, Ba, loss K₂O, Fe₂O₃, Rb</td>
<td>3 %</td>
<td>diagenetic</td>
</tr>
<tr>
<td>chlorite-sericite</td>
<td>musc, chl, py ± calcite ± dio</td>
<td>patchy, alteration of compacted pumice (fiamme), pseudomorphed hbl and px phenocrysts</td>
<td>post–fld pre–to syn–S₁</td>
<td>gain SiO₂, Al₂O₃, Fe₂O₃, MnO, MgO, K₂O, Rb, Ba, loss Na₂O, Sr</td>
<td>4 %</td>
<td>diagenetic</td>
</tr>
<tr>
<td>chlorite-sericite-hematite</td>
<td>chl, ser, hem, ±mag, py ± calcite</td>
<td>patchy or pervasive, interstitial in glomerocrysts, coated fld, stylolites, hbl, filled pelitic fractures</td>
<td>pre–to syn–S₁</td>
<td></td>
<td></td>
<td>diagenetic</td>
</tr>
<tr>
<td>chlorite-epidote</td>
<td>chl, ep, chl, syn–altered, ±calc, ±fld</td>
<td>disseminated patches, replaced hbl, partially replaced fld</td>
<td>post–fld pre–S₁ post–to syn–S₁</td>
<td>gain Al₂O₃, Fe₂O₃, MgO, CaO, Rb loss SiO₂, Na₂O, Sr, Ba</td>
<td>0.67 %</td>
<td>diagenetic-metamorphic</td>
</tr>
<tr>
<td>quartz-sericite</td>
<td>musc, qz, py</td>
<td>pervasive, disseminated</td>
<td>pre–S₁ and pre–to syn–S₁</td>
<td>gain SiO₂, Al₂O₃, K₂O, CaO, Ba, Rb, La, Ce, Nd, los Na₂O, Sr</td>
<td>21 %</td>
<td>hydrothermal</td>
</tr>
<tr>
<td>calcite-quartz-sericite-hematite</td>
<td>ce, qz, hem, ±musc</td>
<td>pervasive, stylolitic</td>
<td>pre–S₁</td>
<td></td>
<td></td>
<td>hydrothermal</td>
</tr>
<tr>
<td>chlorite-pyrite</td>
<td>chl, py</td>
<td>intensive pervasive zones, filled fractures</td>
<td>post–S₁ post–to syn–S₁</td>
<td></td>
<td></td>
<td>hydrothermal</td>
</tr>
<tr>
<td>chlorite-calcite-magnetite</td>
<td>chl, mag, cc</td>
<td>disseminated, replaced phenocrysts</td>
<td>syn–S₁</td>
<td>gain Al₂O₃, CaO, MnO, total Ca²⁺, Rb, loss SiO₂, Na₂O, Sr, Ba</td>
<td>-6.5 %</td>
<td>metamorphic</td>
</tr>
<tr>
<td>sericite</td>
<td>musc</td>
<td>weakly disseminated, defines S₁</td>
<td>syn–S₁</td>
<td></td>
<td></td>
<td>metamorphic</td>
</tr>
<tr>
<td>chlorite-quartz-calcite</td>
<td>chl, qz, py, mag, cc</td>
<td>pressure shadows, hydraulic fractures and tension gashes</td>
<td>post–S₁</td>
<td>gain SiO₂, Al₂O₃, Fe₂O₃, MgO, MnO, CaO, Rb, Sr, loss Na₂O, Ba</td>
<td>4 %</td>
<td>hydrothermal</td>
</tr>
</tbody>
</table>

chlorite-sericite-hematite, chlorite-epidote, quartz-sericite, calcite-quartz-hematite, chlorite-pyrite, chlorite-calcite-magnetite and chlorite-quartz-calcite. The distribution of alteration facies is easily observed in thinsection and handspecimen (Figs. 6.2-6.4) but more difficult to assess on a regional scale due to the patchy nature of the alteration and sparse exposure. Closely spaced drill holes through the northern Central Volcanic Complex east of the Mount Black Fault near the Rosebery ore deposit provide the best information on the distribution of alteration facies (Figs. 6.5 and 6.6).

6.2.1 Alteration geochemistry

Whole-rock geochemical data (Appendix D) were assessed to determine chemical changes associated with the main alteration facies. The Ishikawa et al. (1976) Alteration Index (AI) quantifies Ca and Na depletion and enrichment relative to Mg and K.

\[
AI = 100\left(\frac{K₂O+MgO}{K₂O+MgO+Na₂O+CaO}\right)
\]

The AI for the majority of the samples fall in the range of 25-70, which suggests that the bulk of the northern Central Volcanic Complex is only weakly altered. The AI measures the intensity of
Figure 6.1: Interpretation of the relative timing of alteration assemblages in the northern Central Volcanic Complex east of the Mount Black Fault. S₁ is the bedding-parallel, stylolitic foliation interpreted as a diagenetic compaction and dissolution fabric (Allen, 1990a). S₂ is the regional tectonic cleavage related to Devonian folding. Pre-S₁ sericite is probably a metamorphosed equivalent of early clays. The feldspar-quartz-sericite alteration assemblage may include the growth of early zeolites and the replacement of zeolites and glass by K-feldspar and albite during diagenesis.

plagioclase and glass destruction, but does not discriminate between chlorite and sericite alteration. The AI is also insensitive to carbonate alteration. Strongly carbonate-altered rocks plot in the unaltered range. The Chlorite-Carbonate-Pyrite Index (CCPI) (Large et al., in press a) can be used to partly overcome these problems. The CCPI separates Mg- and Fe-rich chlorites from K-rich sericite.

\[
\text{CCPI} = 100\left(\frac{\text{MgO} + \text{FeO}}{\text{MgO} + \text{FeO} + \text{Na}_2\text{O} + \text{K}_2\text{O}}\right)
\]

Although the CCPI recognises MgCO₃ and FeCO₃, it does not recognise pure calcite (CaCO₃) or rhodochrosite (MnCO₃), which are two of the main types of carbonate associated with hydrothermal alteration around the Rosebery ore deposit (Brathwaite, 1974; Green et al., 1981; Khin Zaw, 1991; Allen, 1997 unpub.). Other carbonates in the northern Central Volcanic Complex include ferron rhodochrosite [(Mn, Fe)CO₃], kutnahorite [CaMn(CO₃)], manganese siderite [Fe(Mn)CO₃] and dolomite [CaMg(CO₃)] (Brathwaite, 1974; Green et al., 1981; Khin Zaw, 1991; Allen, 1997 unpub.). Carbonates containing FeO or MgO will result in CCPI=100, whereas the AI will depend on the Mg/Ca ratio of the carbonate and the presence of other Mg- or Ca-bearing phases (i.e. sericite, chlorite, plagioclase, and epidote) in the alteration assemblage.

Using least-altered samples (defined in Chapter 4, section 4.5.1), least-altered fields on the AI versus CCPI box plot were determined for each of the three compositional groups (established in
Figure 6.2: Alteration facies in rhyolitic pumice-rich rocks from the Kershaw Pumice Formation. Photomicrographs A (ppl), B (xn), C (ppl) and D (xn) are of a feldspar-phyric pumice breccia east of South Hercules (EHP319, 538 m). This breccia is composed of large (6 cm) uncompacted, pink, feldspar-quartz-sericite-altered, fibrous, tube pumice clasts and albite-, sericite- and carbonate-altered plagioclase crystals. The matrix contains disseminated hematite and chlorite-calcite-magnetite. A. Round vesicles (v) are preserved adjacent to the feldspar phenocryst (p). The vesicles are coated with a thin film of sericite. The bubble walls and vesicles consist of albite. B. Well-preserved tube pumice exist at high angles to one another in the pale feldspar-altered domain. Altered plagioclase phenocrysts have thin K-feldspar overgrowths that preserve round vesicles adjacent to the phenocryst. The phenocrysts are dusted with sericite and hematite along the crystal cleavage. The vesicles in C (ppl) and D (xn) are lined with sericite that is partially replaced by fine-grained hematite. The originally glassy walls are composed of feldspar. Feldspar in the vesicles locally preserves radiating textures under crossed nicols. E. Photomicrograph (ppl) of shards in a feldspar-phyric pumice breccia (PR11). The feldspar-quartz-altered shards are coated by sericite and surrounded by fine-grained feldspar-quartz-sericite. Carbonate rhombs overprint the feldspar-altered shards. F. Photomicrograph (ppl) of a perlite fragment and fiamme in a feldspar-phyric pumice-lithic breccia (120R, 122 m). Thin films of sericite define the perlite fractures, whereas the perlitic core is secondary feldspar. The fiamme, a dark chlorite-sericite-rich lens, is overprinted by chlorite-magnetite. G. Photomicrograph (ppl) (EHP319, 538 m) of a chlorite-sericite fiamme enclosed in feldspar-quartz-sericite-altered pumice clasts and shards. Disseminated hematite occurs in the groundmass and as blebs overprinting chlorite-sericite alteration in the fiamme. H. Photomicrograph (ppl) of a chlorite-sericite fiamme with feathery terminations in feldspar-phyric pumice breccia (EHP319, 538 m). Uncompacted pumice clasts are preserved in the feldspar-quartz-sericite domain. Plagioclase crystals are dusted with hematite and sericite.
Figure 6.3: Alteration facies in lavas and sills from the Mount Black Formation. Photomicrographs A (M116, ppl) and B (xn) show fine, arcuate, overlapping perlitic fractures which are preserved by sericite. Feldspar alteration of the perlitic kernels locally preserves radiating fibrous textures. Disseminated hematite exists throughout the groundmass. Photomicrographs C (ppl) and D (xn) are of a flow-banded feldspar-phyric rhyolite lava (R30). Single flow-bands are preserved by feldspar-quartz versus sericite-chlorite ± feldspar alteration. In crossed nicols, fine radiating textures that resemble recrystallised spherulites are observed in the feldspar-rich bands. The original plagioclase phenocrysts (p) are replaced by albite and coated with sericite. Fine-grained, disseminated magnetite exists in the groundmass. E. Photomicrograph (xn) of a densely microspherulitic feldspar-quartz-phyric rhyolite sill (M78). The radiating fibrous structure of the feldspar-quartz-sericite-altered spherulites is preserved by thin films of sericite. F. Photomicrograph (ppl) of quartz-filled lithophysae in a feldspar-phyric dacite (M89). The lithophysae are in a chlorite-rich groundmass. G. Photomicrograph (ppl) of feldspar-hornblende-phyric dacite (M10). The groundmass is densely microspherulitic and the phenocrysts are extensively altered. Hornblende is replaced by chlorite-epidote and plagioclase by albite, sericite and epidote. H. Photomicrograph (ppl) of arcuate perlitic fractures in a feldspar-phyric dacite (PR1). The fractures are filled with sericite, which is partially replaced by chlorite. The glomeroporphyritic cluster of albite-altered plagioclase phenocrysts has interstitial chlorite-magnetite.
Figure 6.4: Alteration facies in coherent and clastic facies from the northern Central Volcanic Complex. A. Photomicrograph (Ppl) of feldspar-phyric pumice breccia from east of South Hercules (EHP319, 538 m). Chlorite-sericite fiamme define the S1 compaction foliation. Uncompacted tube pumice clasts are preserved in the feldspar-quartz-sericite-altered domains. Plagioclase crystals are dusted with hematite, carbonate and sericite. B. Photomicrograph (Ppl) of quartz-sericite-altered feldspar-phyric pumice breccia in the Kershaw Pumice Formation adjacent to the Mount Black Fault (120R, 578 m). Although pervasive quartz-sericite alteration obscures many primary textures, silicified feldspar crystal fragments and sericite-rich fiamme are recognisable. C. Photomicrograph (Ppl) of a vesicular feldspar-phyric basalt clast in a polymictic mafic breccia from the Sterling Valley Volcanics (STP218, 113 m). The clast groundmass comprises fine-grained feldspar, hematite and chlorite. The vesicles are lined with sericite-hematite and filled with chlorite. D. Photomicrograph (Ppl) of a perlitic clast in a feldspar-phyric pumice lithic breccia from the Kershaw Pumice Formation (120R, 226 m). The perlitic fractures are defined by sericite that is partially replaced by chlorite-magnetite. Feldspar alteration of the perlite kernels is incomplete and the centre is altered to chlorite or chlorite-epidote. E. Handspecimen of feldspar-hornblende-phyric dacite (PR57) from the Mount Black Formation, with domainal chlorite-epidote and feldspar-quartz-sericite alteration. Plagioclase phenocrysts are well preserved in the chlorite-epidote-altered domains. F. Photomicrograph (Ppl) of quartz-sericite-altered micropoikilitic dacite (BY2, 53 m). Quartz replaces K-feldspar patches, which enclose laths of albite. The phenocrysts (p) are altered to albite, sericite and carbonate. Calcite rhombs and disseminated chlorite-magnetite overprint the groundmass textures. G. Photomicrograph (xn) of coarse calcite and sericite-hematite styololites in banded carbonate east of Rosebery in the Mount Black Formation (80R, 191 m). H. Photomicrograph (Ppl) of hydraulically brecciated dacite (128R, 62 m). Space-filling chlorite-quartz-carbonate alteration anneals the angular, densely spherulitic dacite fragments.
Figure 6.5: Simplified geological cross section for 900 m south Rosebery (mine grid), showing the main lithofacies and the distribution of the main alteration facies east of the Mount Black Fault. The section line is marked on Figure 3.1 (A-A').
Figure 6.6: Simplified geological cross section for 1320 m north Rosebery (mine grid), showing the main lithofacies and the distribution of the main alteration facies east of the Mount Black Fault. The section line is marked on Figure 3.1 (B-B'). See Figure 6.5 for key to symbols and colours.

section 4.5.3) in the northern Central Volcanic Complex (Fig. 6.7). By comparing the position of data points on the box plot with the least-altered box, the degree of chemical alteration (excluding CaCO$_3$ and MnCO$_3$ alteration) can be estimated. Data points that plot well away from the position of the least-altered box have undergone significant chemical change, whereas points close to the least-altered box generally represent samples that have not.

Alteration effects can also be quantified in terms of mass changes (Barrett and MacLean, 1994). Changes in total mass may be related to dissolution and compaction of originally glassy components, the filling of primary pore spaces, or the replacement of originally glassy or crystalline groundmass by minerals of different density. Mass gains and losses during alteration were calculated using the immobile element method described by MacLean and Barrett (1993), Barrett and MacLean (1994 and 1997). This method assumes that the altered rock was a chemically homogeneous unit (ie. had a single primary
relative mass change = \[(\text{absolute mass change} / \text{component}_\text{last-altered}) \times 100\]

The results are listed in Appendix E. Mass change trends associated with different alteration facies are summarised in Table 6.1 and Figure 6.8. Major absolute mass changes involve gains or losses of greater than 10 g/100 g, small changes 1-10 g/100 g and minor changes <1 g/100 g. Large relative mass changes are >100%, whereas relative mass changes of 50-100% are referred to as moderate.

6.2.2 Sericite alteration
Fine-grained muscovite or sericite is one of the most abundant alteration minerals occurring throughout the northern Central Volcanic Complex. Sericite dusts surfaces on phenocrysts (Fig. 6.3C), vesicles (Fig. 6.2A), clasts, fractures (Fig. 6.3H), spherulites and glass shards (Fig. 6.2E). Sericite is also disseminated throughout the recrystallised groundmasses of coherent units. It has partially to completely replaced plagioclase crystals (Fig. 6.2A) and outlined fibrous feldspar crystals within spherulites (Fig. 6.3E). In pumice-rich units, the surfaces of glass shards and vesicle walls are coated with thin films of sericite (Fig. 6.2A, C and E).

Sericite alteration is associated with variations in whole-rock composition, a major absolute loss of SiO₂ (~10 g/100 g), a smaller loss of Na₂O (~3 g/100 g), minor gains in K₂O, Al₂O₃ and CaO (Fig. 6.8A). These equate to large relative gains in Rb (102%), Ba (230%) and moderate gains in K₂O, P₂O₅, La, Ce, Nd and losses in SiO₂, Na₂O and total mass (Fig. 6.8B).

6.2.3 Carbonate alteration
Carbonate alteration occurs locally in pumice- or scoria-rich rocks of the Kershaw Pumice Formation and Sterling Valley Volcanics. Fine-grained carbonate alteration dusts vesicle walls, fills vesicles and partially replaces the originally glassy walls of pumice or scoria clasts.

6.2.4 Feldspar-quartz-sericite alteration
Feldspar-quartz-sericite alteration facies occurs extensively in the felsic and intermediate rocks of the northern Central Volcanic Complex. The secondary feldspar in the feldspar-quartz-sericite alteration is mainly Na-feldspar (albite) dusted with hematite; however K-feldspar (orthoclase) also occurs locally as relics partially replaced by albite. K-feldspar occurs as overgrowths on primary plagioclase phenocrysts (Fig. 6.2B), annealing fractured crystals and glomerocrysts. Mircopokilitic K-feldspar also occurs in the groundmasses of coherent rhyolites and dacites. Albite formed overgrowths on and also replaced plagioclase phenocrysts (Fig. 6.3C). Sericite exists as flecks within and along the cleavage planes in feldspar crystals, or haloes around albite-altered phenocrysts.

Feldspar-quartz-sericite alteration facies is commonly associated with chlorite-sericite or chlorite-sericite-hematite alteration facies. In pumice-rich units, feldspar-quartz-sericite alteration is dominantal, enclosing dark-green phyllosilicate-rich lenses (Fig. 6.2G and H). The feldspar-quartz-sericite domains preserve relic uncompacted tube pumice. Thin films of sericite define the tube vesicles, whereas the originally glassy walls and vesicles consist of secondary feldspar or a mosaic of feldspar-quartz-sericite (Fig. 6.2B). Locally albite within the vesicles shows a radiating fibrous texture (Fig. 6.2C and D).

In coherent rhyolites and dacites, feldspar-quartz-sericite alteration commonly preserved
Figure 6.8: Mass changes during alteration, calculated using the method of Barrett and MacLean (1994) (Appendix E). A. and B. Absolute and relative mass changes during sericite and feldspar-quartz-sericite alteration. C. and D. Absolute and relative mass changes during chlorite-sericite and chlorite-epidote alteration. E. and F. Absolute and relative mass changes during hydrothermal quartz-sericite alteration. G. and H. Absolute and relative mass changes during chlorite-calcite-magnetite and chlorite-quartz-calcite alteration.
groundmass textures. Originally crystalline (micropoikilitic or microspherulitic) groundmasses are
extensively replaced by feldspar-quartz-sericite (Fig. 6.3E and 6.4F). Originally glassy groundmasses
have been partly altered to fine mosaics of feldspar-quartz-sericite and commonly contain arcuate
perlitic fractures that are filled with sericite or chlorite (Fig. 6.3A and H). Within flow-banded coherent
rhyolites, single, originally spherulitic or crystalline flow-bands have been altered to pink feldspar-
quartz-sericite and originally glassy flow-bands have been altered to green chlorite-sericite (Fig. 6.3C
and D).

Feldspar-quartz-sericite alteration facies is associated with small absolute mass gains in SiO$_2$
($\sim$2 g/100 g), CaO ($\sim$2 g/100 g) and total mass (6 g/100 g), and minor mass gains in Na$_2$O and Al$_2$O$_3$
and minor losses in K$_2$O and Fe$_2$O$_3$ (Fig. 6.8A). Alternatively feldspar-sericite-quartz alteration involved
large relative gains in CaO (140%), moderate relative gains in Na$_2$O, Sr and Ba and losses in MnO,
Fe$_2$O$_3$, K$_2$O, Rb, La, Ce and Nd are also indicated (Fig. 6.8B).

6.2.5 Chlorite-sericite alteration
The chlorite-sericite alteration facies occurs throughout the northern Central Volcanic Complex.
Chlorite-sericite alteration facies commonly enhance or define primary volcanic and compaction
textures. In coherent rhyolites, chlorite-sericite defines fractures, flow-bands and groundmass textures.
In pumice-rich units, dark, bedding-parallel, chlorite-sericite lenses are set in feldspar-rich domains
(Fig. 6.2 F, G, H and 6.4A). Phenocrysts are preserved within these chlorite-sericite lenses but other
primary volcanic textures are mostly destroyed.

Mass changes associated with this alteration style include small absolute gains in SiO$_2$ (2.3 g/
100 g) and K$_2$O (1.2 g/100 g), minor mass gains in Al$_2$O$_3$, MnO and Fe$_2$O$_3$, and a small loss in Na$_2$O
($\sim$2 g/100 g) (Fig. 6.8C). Large relative gains in MnO (190%) and Rb (186%) and moderate gains in
K$_2$O and Rb are also indicated (Fig. 6.8D).

6.2.6 Chlorite-sericite-hematite alteration
Chlorite-sericite-hematite alteration facies occurs in flow-banded coherent rhyolite, monomicitic rhyolite
breccia, monomicitic dacite breccia, pumice-lithic clast-rich breccia and pumice breccia where it defines
a chlorite-sericite-hematite-rich stylolitic fabric (Fig. 4.16).

6.2.7 Chlorite-epidote alteration
Chlorite-epidote alteration typically includes hematite and carbonate and is generally associated with
intermediate and mafic facies in the Sterling Valley Volcanics. Chlorite-epidote also occurs along the
eastern margin of the Mount Black Formation from Tullah to north of the Pieman Road-Murchison
Highway intersection. Chlorite-epidote alteration enveloped and replaced hornblende and partially
altered plagioclase crystals (Fig. 6.3G). In coherent facies, it is interstitial in glomeroporphyritic clusters
and between spherulites, and exists as fine-grained irregular patches in the groundmass. Coherent
basalts have chlorite-epidote-carbonate or chlorite-epidote-hematite-rich groundmasses and epidote-
chlorite-carbonate amygdales. In coherent rhyolites and dacites, some originally glassy groundmasses
are replaced by chlorite-epidote (Fig. 6.4D).

Mass change calculations for chlorite-epidote alteration indicate negligible changes in total mass
(<1 g/100 g), a small absolute gain in CaO (1.2 g/100 g) and minor gains in Al₂O₃, Fe₂O₃, and MgO. Small loss in Na₂O (1.5 g/100 g), and minor losses in K₂O and SiO₂ are indicated (Fig. 6.8E). Alternatively relative mass change calculations show a large gain in MnO (285%) and moderate gains in MgO, CaO and Rb (Fig. 6.8F).

6.2.8 Quartz-sericite alteration
Quartz-sericite alteration is restricted to felsic sills and the volcanic facies adjacent to these sills (Fig. 6.5). Narrow bands of intense quartz-sericite alteration are also associated with major faults and shear zones throughout the sequence. In coherent rhyolites and dacites, quartz-sericite alteration facies comprises pods of quartz enveloped by sericite in the groundmass (Fig. 6.4F). In pumice-rich units, primary plagioclase crystals are silicified and the matrix is pervasively replaced by fine-grained quartz-sericite (Fig. 6.4B).

Quartz-sericite alteration involves a major absolute gain in SiO₂ (~18 g/100 g), smaller gains in Al₂O₃ (1.5 g/100 g) and K₂O (1.2 g/100 g), and a minor loss in Na₂O (Fig. 6.8E). There are large relative gains in La (240%), Ce (218%) and Nd (209%) and moderate lose in Na₂O (Fig. 6.8F). Quartz-sericite alteration also involves considerable increases to the total mass (21%) (Fig. 6.8E) probably as a result of the infilling of porosity.

6.2.9 Calcite-quartz-hematite alteration
Calcite-quartz-hematite alteration occurs in discrete zones in the Kershaw Pumice Formation and the Mount Black Formation, particularly in pumice-lithic clast-rich breccias and monomictic rhyolite breccias at the margins of volcanic lavas (Fig. 3.20). Intense calcite-quartz-hematite alteration masks primary volcanic textures resulting in massive cream and pink rock (Fig. 6.4G). Accessory minerals in the calcite-quartz-hematite alteration may include sericite and pyrite.

6.2.10 Chlorite-pyrite alteration
Chlorite-pyrite alteration is texturally destructive, occurring in discrete zones (up to 10 m across, Fig. 6.5). It also fills both early syn-volcanic fractures and later hydraulic and tectonic fractures.

6.2.11 Chlorite-calcite-magnetite alteration
Chlorite-calcite-magnetite alteration facies is widespread in the felsic Kershaw Pumice Formation and Mount Black Formation. It exists as weak, patchy to pervasive, disseminated fine-grained aggregates in the groundmass or matrix, intersitial in glomeroporphyritic clusters, and between microspherulites and micropoikilitic quartz (Fig. 6.2B, 6.3G, 6.3H and 6.4F). Chlorite-calcite-magnetite replaced primary feldspar and hornblende phenocrysts (Fig. 6.2F), chlorite-sericite and chlorite-sericite-hematite alteration facies, and filled vesicles. Calcite commonly occurs as rhombs that overprint the shard-rich matrices of the pumice-rich units (Fig. 6.2E) and quartzo-feldspathic groundmasses of coherent rhyolites and dacites.

Chlorite-calcite-magnetite alteration is associated with small absolute mass gains in CaO (1.7 g/100 g) and total carbon, and small losses in SiO₂ (~6 g/100 g), Na₂O (2.8 g/100 g) and total mass (6.5 g/100 g), and minor loss in K₂O (Fig. 6.8G). Relative gains are in MnO (350%), CaO (218%), P₂O₅ and Rb and moderate losses in Na₂O and Ba are indicated (Fig. 6.8H).
6.2.12 Chlorite-quartz-calcite alteration

Chlorite-quartz-calcite alteration facies exists in pressure shadows between and adjacent to plagioclase crystals, and fills vesicles, hydraulic fractures (Fig. 6.4H) and extension fractures in all rock types. Chlorite-quartz-calcite alteration facies is associated with tourmaline in some faults (Fig. 6.5).

Mass-change calculations indicate small absolute mass gains in SiO₂ (2 g/100 g), CaO (2 g/100 g), minor gains in Al₂O₃ and total carbon, small losses of Na₂O (~3 g/100 g) (Fig. 6.8G). Alternatively, mass change calculations indicate large relative gains in MnO (264%) and CaO (319%) and moderate gains in P₂O₅ (Fig. 6.8H).

6.3 Early foliations in pumice-rich facies

The compaction foliation (S₁) is a pre-tectonic foliation that is defined by the bedding-parallel alignment of sericite or chlorite-sericite lenses (fiamme) and chlorite-sericite-hematite stylolites in pumiceous units (Chapter 4).

6.3.1 Sericite and chlorite-sericite fiamme

The term “fiamme” is used to describe the textural appearance of the sericite and chlorite-sericite lenses and does not imply any particular origin. Sericite and chlorite-sericite fiamme in pumice-rich facies have a wide variety of sizes (1 mm to 20 cm), length to height ratios (up to 12:1) and shapes including: flame-like, bowtie, irregular branching and blocky shapes (Fig. 6.9). They are commonly as large or larger than the maximum uncompacted pumice clast size in the pumice-rich facies. They have either feathery terminations or end in long thin wispy chlorite-sericite or chlorite-sericite-hematite stylolites (Fig. 6.9). Sericite and chlorite-sericite fiamme are aligned parallel to bedding and often thin between non-vesicular lithic clasts and plagioclase crystals. Internally the sericite and chlorite-sericite fiamme can be aphyric or plagioclase-phyric, massive, foliated, contain chlorite-sericite-hematite stylolites, or rarely preserve delicate feldspar-filled tube vesicles (Fig. 6.9). Chlorite-sericite-hematite stylolites anastomose along the length of the fiamme, commonly branching and deviating around plagioclase phenocrysts, and extend past the end of the fiamme into the matrix (Fig. 6.9).

6.3.2 Chlorite-sericite-hematite stylolites

Chlorite-sericite-hematite stylolites in pumice-rich facies are irregular, anastomosing sutures characterised by the concentration of fine-grained hematite (Fig. 4.4). They occur along the margin of originally glassy clasts, along tube vesicle walls in pumice clasts and in the matrix of pumice-rich facies (Figs. 4.4 and 6.9).

6.3.3 Interpretation of early foliations in pumice-rich facies

The sericite or chlorite-sericite fiamme have delicate feathery terminations and locally preserve flattened tube pumice textures indicating that they are altered compacted pumice clasts. The range in textures from uncompacted feldspar-quartz-sericite-altered tube pumice clasts to sericite or chlorite-sericite fiamme with feldspar-filled tube vesicles to massive sericite or chlorite-sericite fiamme, corresponds with increasing length to height ratios. This suggests that tube pumice clasts have undergone varying degrees of compaction. Fiamme shapes, textures and sizes indicate that they may comprise a single flattened pumice clast, part of a pumice clast or aggregates of pumice clasts. Blocky fiamme are
Figure 6.9: Characteristic shapes and textures of fiamme in the northern Central Volcanic Complex. Sketches are from samples of pumice breccia which contain limited evidence of deformation related to the regional tectonic cleavage (S2). A. Plagioclase porphyritic, flame-like chlorite-sericite fiamme. B. Chlorite-sericite fiamme with long wispy stylolitic ends. This fiamme contains chlorite-sericite-hematite stylolites. C. Bowtie-shaped chlorite-sericite fiamme with sparse feldspar-filled tube vesicles and chlorite-sericite-hematite stylolites developed parallel to the tube vesicles. D. Flame-like, branching sericite fiamme with chlorite-sericite or chlorite-sericite-hematite stylolites. E. Ragged fiamme that shows variable compaction of a pumice clast or clasts. The left-hand end is chlorite-sericite altered and contains uncompacted feldspar-filled tube vesicles. The right-hand end is massive chlorite-sericite and contains chlorite-sericite stylolites parallel to the length of the fiamme. F. Chlorite-sericite fiamme with a brush-like, poorly compacted end containing feldspar-filled vesicles. G. Chlorite-sericite fiamme composed of uncompacted, altered pumice clast. The plagioclase phenocryst is albite-altered, tube vesicles have been filled with feldspar-quartz-sericite and the originally glassy vesicle walls have been replaced by chlorite-sericite. H. Blocky fiamme that comprises a single chlorite-sericite altered perlite clast. Chlorite-sericite-hematite stylolites are developed at the margin of the clast.
interpreted to be the product of sericite or chlorite-sericite alteration of originally glassy non-vesicular lithic clasts and show no evidence of flattening (cf. Allen, 1988). Randomly oriented and undeformed feldspar-quartz-sericite-altered pumice clasts and shards in the pumice-rich facies indicate that compaction did not involve hot welding and plastic deformation (cf. Smith, 1960). In pumice-rich facies, the pressure of burial can flatten long tube structures, particularly if pumice clasts have previously been altered to mechanically weak minerals, and squeeze clasts together to produce a compact rock that superficially resembles a welded pyroclastic deposit (Fiske, 1969; Schmincke et al., 1982; Branney and Sparks, 1990). In the northern Central Volcanic Complex, compaction of pumice clasts to produce fiamme is interpreted to have occurred prior to regional deformation (Chapter 4) as a result of overburden pressure (cf. Allen and Cas, 1990 unpub.; Branney and Sparks, 1990; Taylor et al., 1990).

Although sericite and chlorite-sericite fiamme occur dominantly in pumice-rich facies they also occur in some coherent rhyolites and dacites where they are interpreted to represent the flattening of originally pumiceous flow-bands either as a result of shear during flowage (cf. Pichler, 1981 in Branney and Sparks, 1990) or as a result of burial compaction.

The chlorite-sericite-hematite stylolitic foliation (S1) is interpreted as a diagenetic compaction and dissolution fabric that formed by the dissolution of soluble components, particularly glass, as a result of pressure during burial (Allen, 1990a unpub.; Allen and Cas, 1990 unpub.).

6.4 Overprinting relationships and timing of alteration facies

The relative timing of alteration facies was determined by documenting overprinting relationships among the alteration mineral assemblages, the compaction foliation (S1) and regional tectonic cleavage (S2) (Fig. 6.1).

Thin films of sericite that line vesicle walls in pumice-rich and vesicular facies (Fig. 6.2A and C) pre-date the infilling of vesicles and compaction. This alteration facies is overprinted by all other alteration facies.

In pumice-rich units, domains of feldspar-quartz-sericite alteration overprint the sericite films within vesicles (Fig. 6.2C). The feldspar-quartz-sericite alteration fills uncompacted vesicles and replaces the originally glassy vesicle walls in the pumice clasts. This indicates that the feldspar-quartz-sericite alteration occurred prior to compaction and deformation.

Carbonate-filled vesicles in altered pumice clasts lack sericite films, which suggests that carbonate alteration began very early, synchronous with the infilling of the primary porosity and prior to compaction. Massive calcite-quartz-hematite alteration locally contains relic feldspar-quartz-sericite altered clasts and bedding-parallel stylolites. This suggests that calcite-quartz-hematite alteration post-dates feldspar-quartz-sericite alteration but was pre- to syn-compaction.

Chlorite-sericite has partially replaced sericite rims on originally glassy surfaces, and altered compacted pumice clasts or fiamme that define the S1 compaction foliation. This implies that the chlorite-sericite alteration assemblage is pre-to syn-S1. Some pumice clasts have been partially altered
to both chlorite-sericite and feldspar-quartz-sericite and variably compacted, suggesting that chlorite-
sericite alteration is synchronous with or post-dates feldspar-quartz-sericite alteration. Altered perlite
with a sequence of sericite, feldspar-quartz-sericite and chlorite-sericite alteration from fracture to
core, also suggests that chlorite-sericite alteration post-dates feldspar-quartz-sericite.

The chlorite-sericite-hematite stylolitic fabric defines the S₁ foliation and is crenulated by the
regional cleavage (S₂) (Fig. 4.4) implying that it is syn-S₁.

Chlorite-epidote alteration overprints the early sericite coating of glass surfaces and chlorite-
sericite fiamme. It is interpreted to post-date the S₁ compaction fabric and is pre- to syn-S₂.

There are two periods of quartz-sericite alteration (Fig. 6.1). Quartz-sericite alteration prior to
compaction and S₁ filled uncompacted vesicles in pumice clasts. Quartz-sericite alteration is also
associated with shear zones and faults parallel to S₂.

Chlorite-calcite-magnetite alteration facies replaces and post-dates chlorite-sericite fiamme (Fig.
6.2F) and chlorite-sericite-hematite stylolites that define S₁, and are commonly aligned parallel to S₂
cleavage and associated with syn-S₂ veins.

Chlorite-pyrite alteration is associated with shear zones parallel to S₂ and undeformed post-S₂
brittle fractures and faults.

Minerals in the chlorite-quartz-calcite alteration facies show no evidence of deformation and
are not aligned parallel to the regional foliations (S₁ or S₂). This facies overprints sericite, epidote and
chlorite-rich alteration facies and is interpreted to post-date S₂.

6.5 Interpretation of alteration facies in the northern Central Volcanic Complex

6.5.1 Textural and chemical modification of volcanic facies

Abundant spherulites, lithophysae, micropoikilitic texture and relic perlite indicate that volcanic rocks
in the northern Central Volcanic Complex were initially partly crystalline and partly glassy. Alteration
processes have modified both the texture and chemistry of these glassy and crystalline volcanic rocks.
The alteration processes were associated with diagenesis, hydrothermal alteration, metamorphism and
deformation. In addition, originally glassy domains were thermodynamically unstable and thus were
readily susceptible to hydration and devitrification (Fisher and Schmincke, 1984; Friedman and Long,
1984).

Hydration of glass involves the absorption of water into solid glass and the modification of the
glass structure, either during cooling or at ambient temperatures (Ross and Smith, 1955; Friedman and
Long, 1984). Perlite is traditionally interpreted to have formed due to the release of stress during
volume changes associated with hydration (Ross and Smith, 1955; Drysdale, 1991). Compositional
changes accompanying hydration include minor losses in silica and alkalis (Noble, 1967; Friedman and
Long, 1984; Mungall and Martin, 1994).
Devitrification refers to the nucleation of crystals in solid glass below the glass transition temperature (Lofgren, 1971). Devitrification results in the gradual conversion of glass to a fine-grained granular crystalline aggregate (Bigger and Hanson, 1992; McArthur et al., 1998) and may be accompanied by changes in whole-rock composition (Lipman, 1965; Lofgren, 1971; Friedman and Long, 1984).

Hydrothermal alteration results from the interaction of rock with hot water or steam during changing thermal and chemical conditions (Henley and Ellis, 1983). Hydrothermal alteration involves ion-exchange reactions, phase transformations, dissolution and precipitation of new minerals. Hydrothermal alteration results in significant changes in texture, mineralogy and whole-rock geochemistry.

Diagenesis encompasses the mineral changes that occur in response to the regional geothermal gradient in the depositional basin (Torres et al., 1995). During diagenesis of glassy volcanic facies, significant textural and mineralogical changes can be produced by precipitation of cement, dissolution and replacement of original components, especially glass, and compaction (Fisher and Schmincke, 1984; Marsaglia and Tazaki, 1992; Torres et al., 1995). The transitions between diagenesis and metamorphism, and diagenesis and hydrothermal alteration have not been rigorously defined (Pettijohn et al., 1972; Fisher and Schmincke, 1984; Morrow and McIlreath, 1990).

Regional metamorphism involves pervasive isochemical mineralogical and textural changes in response to increasing pressure and temperature (Yardley, 1989). Regional metamorphic assemblages overprint pre-existing diagenetic and hydrothermal textures and mineralogies, however they generally do not involve significant changes in chemical composition. The northern Central Volcanic Complex has undergone regional metamorphism to lower greenschist facies contemporaneous with the Devonian deformation (Green et al., 1981; Walshe and Solomon, 1981; Corbett and Solomon, 1989; Corbett, 1992). Greenschist facies metamorphic assemblages include minerals (e.g. sericite, epidote and carbonate) that are also common in diagenetic and hydrothermal alteration assemblages. Deformation on both local and regional scales may modify or destroy original volcanic textures and create new fabrics.

In this study diagenetic, metamorphic and hydrothermal alteration facies have been distinguished on the basis of timing, texture, mineralogy, intensity and distribution. Diagenetic facies are regional in distribution and pre-date, or are synchronous with, the stylolitic S1 compaction foliation (Fig. 6.1). Diagenetic alteration is generally weak, with volcanic textures and plagioclase crystals remaining well preserved. Regional metamorphic assemblages are regional in extent, but post-date S1 and are syn-to post-S2. They are typically weak and preserve or modify primary textures. Alteration facies that are local in distribution are interpreted to be related to discrete hydrothermal cells or the cooling of sills. Hydrothermal alteration can also be distinguished geochemically from diagenetic and metamorphic alteration using the Al (Ishikawa et al., 1976) and CCPI (Large et al., in press a). Hydrothermal alteration is moderate to intense, locally pervasive, transgressing facies and stratigraphic boundaries, modifies primary plagioclase crystals and may destroy primary textures.

Some difficulty exists in differentiating between diagenetic and hydrothermal alteration facies at the margins of hydrothermal alteration zones, as the conditions of hydrothermal alteration in submarine
volcanic successions overlap with those of diagenetic alteration (Iijima, 1974; Henneberger and Browne, 1988; Allen, 1997 unpub.). In addition, the diagenetic and some hydrothermal mineral assemblages have been recrystallised and variably replaced by regional metamorphic mineral assemblages. However, the original assemblages can be inferred using local relic textures and by comparison with younger hydrothermally and diagenetically altered volcanic successions.

6.5.2 Diagenetic alteration

Weak sericite, carbonate, feldspar-quartz-sericite, chlorite-sericite and chlorite-sericite-hematite alteration facies coated original surfaces, infilled primary porosity and replaced glass in the northern Central Volcanic Complex prior to or synchronous with compaction. These alteration facies are early (pre-to syn-compaction) and regionally extensive, and are thus considered to be diagenetic facies or metamorphic equivalents of diagenetic facies.

The coating of original surfaces

The first stage in the alteration of the northern Volcanic Complex is represented by sericite, chlorite-sericite-hematite and carbonate films lining vesicles, along fractures and coating glassy clasts (Fig. 6.2A, 6.3H and 6.4C). In felsic volcaniclastic units of the Kershaw Pumice Formation, thin films of sericite coat vesicles, originally glassy shards and crystals. This is texturally identical to the first stage of diagenetic alteration of felsic pumice and glass shards where thin films of smectite clay and opal develop on all glass surfaces (Iijima, 1974; Walton, 1975; Henneberger and Browne, 1988; Tsolias-Katagas and Katagas, 1989; Noh and Boles, 1989; Marsaglia and Tazaki, 1992; Torres et al., 1995). Thus sericite may be the metamorphic equivalent of an early clay alteration which pre-dates sealing of the porosity and compaction, as suggested by Allen (1990a unpub.) and Allen and Cas (1990 unpub.).

Sericite filling perlitic fractures in coherent rhyolites and dacites of the Kershaw Pumice Formation and Mount Black Formation (Fig. 6.3A, 6.3H and 6.4D) could also be replacing early clays produced by hydration of volcanic glass or diagenetic precipitation (Lipman et al., 1969; Boles and Coombs, 1977; O'Keefe, 1984; Noh and Boles, 1989). Sericite that replaced glass shards and pumice suggests that early alteration involved the conversion of glass to clays (cf. Ratterman and Surdam, 1981; Marsaglia and Tazaki, 1992; Tazaki and Fyfe, 1992).

In the felsic Kershaw Pumice Formation and Mount Black Formation, this early diagenetic alteration is associated with minor increases in $K_2O$ and $Al_2O_3$ (Fig. 6.8A). Similar trends result from the conversion of silicic glass to clay (Noh and Boles, 1989; Passaglia et al., 1995).

In mafic volcaniclastic facies of the Sterling Valley Volcanics, chlorite-sericite-hematite facies coated vesicles, crystals, clasts and fractures (Fig. 6.4C). This is texturally similar to the first stage of diagenesis of basalts, involving palagonitisation of basaltic glass, initially on glass surfaces, along fractures and around vesicles (Zhou and Fyfe, 1989). Fine Fe- and Ti-oxides are commonly associated with palagonitisation (Dimroth and Lichtblau, 1979). This is consistent with the presence of hematite in this early alteration facies.

Carbonate is also present as thin films on vesicle walls, infilling vesicles, and partially replaced plagioclase crystals in felsic facies in the Kershaw Pumice Formation. This suggests that carbonate
alteration began early in the diagenetic history. It is possible that carbonate growth continued synchronous with feldspar-quartz-sericite alteration as Ca-rich carbonates consume calcium released during albitisation (Boles, 1982; Marsaglia and Tazaki, 1992).

**Pore-filling feldspar-rich alteration**

The primary porosity has been filled by feldspar-quartz-sericite and phyllosilicate-rich alteration facies (chlorite-sericite and chlorite-sericite-hematite). This alteration must have followed the coating of original surfaces by clays or palagonite.

In pumice-rich units, K-feldspar alteration nucleated on plagioclase phenocrysts and spread outwards filling pore space and replacing the glassy bubble walls (Fig. 6.2A). The K-feldspar could be the original diagenetic phase or else, it replaces a diagenetic phase such as zeolites. Radiating fibrous textures in the feldspar are locally preserved in vesicles within the pumice breccias and in the originally glassy kernels of petrified dacites (Fig. 6.2C, 6.2D, 6.3A, 6.3B, 6.10A and 6.10B). In some young volcanic successions, fibrous radiating zeolites fill vesicles (Fig. 6.10) and replace glass shards and glassy kernels in perlite (Dimroth and Lichtblau, 1979; Tsolis-Katagas and Katagas, 1989; Noh and Boles, 1989;

![Image of alteration facies filling vesicles in pumice clasts](image)

**Figure 6.10:** Alteration facies filling vesicles in pumice clasts from the northern Central Volcanic Complex and the Miocene Green Tuff Belt, Central Hokuroku Basin, Japan. A. Photomicrograph (ppl) of vesicles in a pumice clast from pumice breccia east of South Hercules (EHP319, 538 m). The vesicles are lined with sericite and filled with feldspar. The originally glassy walls are composed of feldspar. B. Photomicrograph (xn) of feldspar in the vesicles locally preserving radiating textures. C. Photomicrograph (ppl) of mordenite filled vesicles in pumice breccia from the Miocene Green Tuff Belt, Japan (16 395 m). D. Photomicrograph (ppl) of vesicles lined with chlorite and clays, and filled with fibrous mordenite (16 395 m).
This suggests that zeolites filled primary pore spaces and replaced shards and felsic glass early in the diagenetic history of the northern Central Volcanic Complex. As the temperature and pressure increased, secondary feldspar (K or Na) replaced early zeolites. Subsequent recrystallisation during metamorphism destroyed much of the fibrous textures.

The extensive replacement of K-feldspar by albite in the Kershaw Pumice Formation and Mount Black Formation, suggests that there were at least two stages of feldspar alteration. Munha et al. (1980) suggested that at lower temperatures (<150°C), Na⁺ in glass can be exchanged for K⁺ in seawater, resulting in precipitation of K-feldspar and K-smectites. However, K-feldspar has not been reported as a direct product of glass alteration (Surdam and Sheppard, 1978). An intermediate phase between glass and K-feldspar is proposed. Iijima and Hay (1968), Surdam and Sheppard (1978) and Hay and Guldman (1987) recognised the replacement of zeolites and analcime by K-feldspar. The replacement of K-feldspar by albite could reflect the exchange of K⁺ in the rock with Na⁺ in seawater at greater depths and temperatures (100-150°C) (Munha et al., 1980; Boles, 1982).

Feldspar-quartz-sericite alteration where albite has replaced K-feldspar in the Kershaw Pumice Formation and Mount Black Formation, is associated with a gains in CaO, SiO₂, Na₂O, Sr and Ba, consistent with seafloor albitionation (Boles and Coombs, 1977; Boles, 1982).

**Dissolution and phyllosilicate alteration of glass**

Phyllosilicate-rich alteration facies (chlorite-sericite and chlorite-sericite-hematite) occur in originally glassy shards, glassy bubble walls in pumice clasts (Fig. 6.2F), glassy flow bands (Fig. 6.3C), and the interstitial glass between spherulites and lithophysae (Fig. 6.3F). This is consistent with the phyllosilicate-rich alteration facies being the metamorphosed equivalent of clays and chlorite that replaced glass during diagenesis (cf. Iijima, 1974; Hay and Guldman, 1987; Noh and Boles, 1989). The presence of a spaced chlorite-sericite-hematite stylolitic dissolution foliation in phyllosilicate-altered domains in pumice breccias suggests that original glass was dissolved and that clays and Fe-oxides precipitated synchronous with compaction.

In phyllosilicate-altered felsic rocks, gains in MgO are consistent with the formation of smectite, chlorite and other Mg-silicates during diagenesis (cf. Hajash and Chandler, 1981; Shiraki and Iiyama, 1990). Losses in Na₂O are consistent with the leaching of alkalis into seawater. The abundance of hematite may reflect the oxidation of Fe³⁺ during alteration of glass to clays (eg. Klein and Lee, 1984).

**6.5.3 Alteration transitional between diagenesis and metamorphism**

In the northern Central Volcanic Complex, chlorite-epidote alteration pre-dates regional deformation and metamorphism, existing in originally glassy domains and replacing chlorite-sericite-rich facies. Negligible absolute and total mass changes associated with chlorite-epidote alteration suggest that chlorite-epidote grew in response to the regional geothermal gradient late in the diagenetic history. Chlorite and epidote have been observed as the direct alteration product of dacitic to basaltic glass and clays at relatively shallow depths (420 m) in modern geothermal regions (White and Sigvaldson, 1962). They are also typical of low temperature metamorphism and the growth of these minerals may bridge the boundary between diagenesis and metamorphism where glass and diagenetic clays are replaced by phyllosilicates (cf. Baragar et al., 1979).
The occurrence of chlorite-epidote alteration facies in rocks in the core of the regional anticline between Tullah and the Pieman Road-Murchison Highway intersection suggests that this alteration facies is associated with the deepest stratigraphic level in the northern Central Volcanic Complex. This is consistent with a high-grade diagenetic alteration zone developed at depth in response to increased geothermal gradient (cf. Iijima, 1974; Hay, 1978).

6.5.4 Hydrothermal alteration
Quartz-sericite, calcite-quartz-hematite, chlorite-pyrite, chlorite-calcite-magnetite and chlorite-quartz-calcite alteration in the northern Central Volcanic Complex east of the Mount Black Fault are limited in extent, cross cut stratigraphic boundaries (Fig. 6.5 and 6.6), are texturally pervasive, have high alteration indices and show significant geochemical deviations from the least-altered signature (Fig. 6.8). These characteristics are consistent with the products of hydrothermal alteration rather than more regional diagenetic alteration or metamorphism. The hydrothermal alteration facies are commonly associated with faults, shear zones and syn-volcanic intrusions (Fig. 6.5 and 6.6). Some quartz-sericite altered rocks adjacent to the Mount Black Fault and Rosebery, have alteration indices consistent with VHMS-style alteration (Fig. 6.7A).

The earliest hydrothermal activity generated quartz-sericite alteration facies within syn-volcanic felsic intrusions and in the adjacent volcanic rocks (Fig. 6.5). This is characterised by SiO₂ enrichment an order of magnitude greater than the mass changes associated with the diagenetic alteration (Fig. 6.8). Seawater and pore fluids, heated by the syn-volcanic intrusions, circulated through the intrusions leaching SiO₂ and other major elements from volcanic glass. As the circulating fluids cooled at the margins of the intrusions, the SiO₂ and K₂O were precipitated from solution (Sigurdsson, 1977; Hanson and Wilson, 1993). Hydrothermal activity also generated calcite-quartz-hematite alteration. Intense calcite-quartz-hematite alteration occurs in a series of carbonate lenses east of Rosebery (Fig. 6.4G and 6.6) and C-O isotope signatures of the calcite indicated that Cambrian hydrothermal fluids were responsible for this massive, texturally destructive alteration (Chapter 3).

Chlorite-pyrite and chlorite-calcite-magnetite alteration facies are associated with syn-S2 shear zones and brittle fractures and may be related to hydrothermal activity accompanying Devonian deformation and granite intrusion. The youngest hydrothermal product in the northern Central Volcanic Complex is the chlorite-quartz-calcite assemblage that post-dates regional deformation and the formation of S2.

6.5.5 Diagenetic alteration zones
Two regionally developed alteration zones (albite zone and epidote zone) are recognised in the northern Central Volcanic Complex. The albite zone is characterised by feldspar-quartz-sericite and chlorite-sericite alteration facies. It is thick (>2 km) and encompasses the Hercules Pumice Formation, the Kershaw Pumice Formation and most of the Mount Black Formation.

The epidote zone is characterised by chlorite-epidote, chlorite-sericite-hematite and feldspar-quartz-sericite alteration facies. The epidote zone is less extensive than the albite zone and occurs in the Mount Black Formation and Sterling Valley Volcanics at the stratigraphic base of the northern Central Volcanic Complex.
The change from albite zone to epidote zone with stratigraphic depth is consistent with diagenetic alteration zonation (cf. Iijima, 1974, 1978). Because diagenetic alteration develops in response to increasing temperature with depth during burial it forms in a sequence of flat lying alteration zones (Fig. 6.11B) (Iijima, 1974; 1978; Fisher and Schmincke, 1984; Utada, 1991). Alteration zones are characterised by mineral assemblages which reflect the reaction of glass and diagenetic minerals with interstitial seawater at a particular temperature (anywhere between 0° and 250°C) (Iijima, 1978; Hay, 1978; Utada, 1991). With increasing temperature glass is initially altered to clays and zeolites and then

A. Northern Central Volcanic Complex

B. Central Hokuroku Basin

C. Rosebery K-lens VHMS deposit

to albite, chlorite, epidote and quartz (Fig. 6.12). In the northern Central Volcanic Complex, thick (> 1 km) diagenetic alteration zones with high-temperature assemblages (albite-quartz-sericite and chlorite-epidote) suggest that they developed in response to a high-grade diagenetic alteration system that involved an elevated geothermal gradient (cf. Utada, 1991).

The vertical zonation of diagenetic alteration in the northern Central Volcanic Complex contrasts with models for hydrothermal alteration haloes associated with VHMS mineralisation (Fig. 6.11). Hydrothermal alteration haloes are interpreted to have formed as a result of the interaction of hot (>300°C) hydrothermal fluids with volcanic facies (Gemmell and Large, 1992; Large, 1992). The migration of hydrothermal fluids and mixing with cold seawater at the margins of the hydrothermal system produces 3-dimensional alteration zones composed of successively lower temperature mineral

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assemblages away from the site of mineralisation (Urabe et al., 1983; Date et al., 1983; Large, 1992). Alteration haloes associated with Australian VHMS deposits commonly include: a quartz-sericite zone, chlorite zone, sericite zone and carbonate zone (Fig. 6.11C) (Gemmell and Large, 1992; Large, 1992; Doyle, in press; Large et al., in press b).

### 6.5.6 Controls on the distribution of diagenetic alteration facies

Although diagenetic alteration facies in the northern Central Volcanic Complex are widespread, their distribution is locally patchy (Fig. 6.5 and 6.6). The patchy distribution and variation in intensity of the diagenetic alteration facies can be related to the original volcanic facies, especially its composition, the presence of volcanic glass, variations in porosity and permeability and external conditions such as pressure, temperature, fluid composition and fluid/rock ratio. In areas of intense alteration, the original volcanic facies had less influence on the distribution and mineralogy of the alteration products.

During weak diagenetic alteration, the primary volcanic composition has strongly influenced the alteration assemblage, the distribution of these assemblages and the textures formed. The alteration assemblages which dominate in felsic rocks (mainly the Hercules Pumice, Kershaw Pumice and Mount Black Formations) are sericite, feldspar-quartz-sericite, chlorite-sericite, chlorite-sericite-hematite and carbonate whereas the more mafic Sterling Valley Volcanics are typically dominated by chlorite-sericite, chlorite-sericite-hematite, chlorite-epidote and carbonate-rich alteration assemblages. Similarly, Tsolias-Katagas and Katagas (1989) found that original compositional differences between rhyolitic and dacitic lithic fragments in the Santorini tuff controlled the mineralogy of diagenetic zeolites.

The original proportion and distribution of glassy versus crystalline domains were also important (cf. Lee and Klein, 1986). Relic textures indicate that prior to alteration, flow-banded rhyolites in the Kershaw Pumice Formation and the Mount Black Formation comprised alternating glassy and crystalline spherulitic or micropoikilitic bands. The crystalline flow-bands and spherulites have been recrystallised and replaced by feldspar-quartz-sericite (Fig. 6.3C and D). In contrast, the glassy flow-bands and glassy domains between spherulites have had a more complex history. Typically, glass surfaces were coated by clays (subsequently metamorphosed to sericite) and glassy domains were replaced by phyllosilicate-rich assemblages of clays, chlorite-sericite, chlorite-sericite-hematite and chlorite-epidote. Later phyllosilicate assemblages overprint earlier assemblages in the originally glassy domains.

The porosity and permeability of volcanic facies within the northern Central Volcanic Complex would have varied considerably over short distances, due to both the distribution of primary volcanic textures and post-depositional modification. Volcaniclastic facies, particularly pumice breccias and pumice-lithic clast-rich breccias, would have initially had a very high porosity, which was subsequently filled by feldspar-quartz-sericite alteration and reduced by compaction. In contrast, the coherent facies of lavas and sills had a much lower porosity and permeability, which were controlled by fractures produced by quenching, flowage and hydration. Alteration in the coherent facies of lavas and sills has progressed as fronts moving outward from fractures into the less altered domains. During early alteration of perlite, the perliteic cracks behaved as fluid pathways. Fractures were sealed with clays (replaced by sericite), closing the primary porosity. Subsequent feldspar-quartz-sericite or feldspar-quartz-sericite after diagenetic zeolites replaced the remaining glassy perlite cores. This suggests that the glassy core was more permeable or more readily altered than the clay-filled fractures.
In volcaniclastic facies, the inter- and intra-particle pore space controlled the porosity and permeability. In the northern Central Volcanic Complex, grain size, type and sorting were all important in influencing the distribution of early alteration facies. Early feldspar-quartz-sericite and chlorite-sericite alteration in well-sorted pumice breccia, although commonly patchy, was pervasive and poorly focused. In poorly sorted volcaniclastic facies, such as pumice-lithic clast-rich breccia and polymictic mafic breccia, the porosity and permeability were initially much more variable and diagenetic alteration facies in these rocks typically have complex distributions.

6.6 Comparison of diagenetic and VHMS-related hydrothermal alteration

The products of diagenesis and hydrothermal alteration can be discriminated on the basis of distribution, mineralogy, timing, degree of textural modification, and degree of chemical modification as reflected by alteration indices and mass-change calculations. In this section, diagenetic alteration facies in the northern Central Volcanic Complex east of the Mount Black Fault are compared with footwall hydrothermal alteration facies associated with mineralisation at Rosebery (Table 6.2). The host stratigraphy at Rosebery is summarised in Chapter 2. The mineralisation and alteration at Rosebery have been described elsewhere by Green et al. (1981), Green and Iff (1989), Allen and Cas (1990 unpub.), Lees et al. (1990) and Allen (1994b unpub.). The Rosebery deposit comprises sixteen separate stacked ore lenses that have overlapping alteration haloes with characteristic mineralogy and geochemistry (Large et al., in press b). These alteration haloes comprise a series of zones passing away from the deposit: quartz-sericite, Mn-carbonate, sericite and chlorite (Lees et al., 1990; Large et al., in press b). Unlike many other VHMS deposits, no chlorite alteration pipe underlies Rosebery, although limited stratabound chlorite occurs below the copper-rich ore lenses in the south end of the mine (Brathwaite, 1974; Green et al., 1981).

Diagenetic alteration in the northern Central Volcanic Complex is generally weak to moderate in intensity, and original groundmass and matrix textures and feldspar phenocrysts are well preserved. In fact, some diagenetic changes actually enhanced primary volcanic textures. For example, diagenetic sericite alteration has enhanced perlitic fractures, and outlined shards and pumice clasts; feldspar-quartz alteration has preserved uncompacted shards and pumice clasts. Diagenetic chlorite-rich assemblages were texturally more destructive, feldspar phenocrysts are the only recognisable original component in the chlorite-sericite fiamme. More intense diagenetic alteration was pervasive and destroyed primary groundmass textures, but feldspar phenocrysts are widely preserved.

In contrast, footwall hydrothermal alteration at Rosebery is generally intense and totally texturally destructive close to the massive sulfide lenses (Lees et al., 1990). All primary volcanic textures including the feldspar phenocrysts have been destroyed by quartz-sericite and sericite-chlorite alteration. In areas of less intense hydrothermal alteration, the distribution of alteration facies is not always pervasive, feldspar phenocrysts have been significantly altered (Lees et al., 1990) and groundmass textures have been modified and/or destroyed by carbonate, sericite-quartz and sericite-chlorite alteration (Large et al., in press b). Pumice and shards textures are well preserved in patchy carbonate alteration (Allen, 1997 unpub.). The effects of weak hydrothermal alteration are difficult to distinguish from the effects of regional diagenetic alteration. Weak sericite alteration formed disseminated grains in the groundmass
Table 6.2: Characteristics of diagenetic alteration facies in the northern Central Volcanic Complex (this study) and footwall hydrothermal alteration facies at Rosebery (this study; Allen, 1997 and Large et al., 2001b). Mass change calculations in Appendices E, F, G and H. AI= alteration index (Ishikawa et al., 1976), fld=feldspar, ser=sericite, chl=chlorite, CO3=carbonate.

<table>
<thead>
<tr>
<th>Alteration facies</th>
<th>Alteration features</th>
<th>Volcanic recency</th>
<th>Phenocrysts</th>
<th>AI</th>
<th>Total mass change</th>
</tr>
</thead>
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<tr>
<td><strong>DIAGENETIC</strong></td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>sericite</td>
<td>pervasive alteration of glass surfaces and primary volcanic fractures</td>
<td>enhanced</td>
<td>dusted with ser</td>
<td>20-40</td>
<td>-9 %</td>
</tr>
<tr>
<td>carbonate</td>
<td>covered originally glassy surfaces</td>
<td>preserved except in intense alteration</td>
<td>preserved</td>
<td>35-45</td>
<td></td>
</tr>
<tr>
<td>feldspar-quadrate-sericite</td>
<td>pervasive alteration of groundmasses and selective or patchy alteration</td>
<td>preserved except in intense alteration</td>
<td>replaced by secondary ser, CO3-chlorite alteration</td>
<td>Na-K-30-45</td>
<td>-1 to +5 %</td>
</tr>
<tr>
<td>chlorite-sericite</td>
<td>chlorite-sericite-hematite</td>
<td>patchy to pervasive, disseminated, interstitial to crystals, stylolites</td>
<td>destroyed to moderately preserved</td>
<td>CO3-chlorite alteration preserved</td>
<td>45-50</td>
</tr>
<tr>
<td>chlorite-epidote</td>
<td>pervasive alteration of groundmass</td>
<td>preserved</td>
<td>chl-ser alteration</td>
<td>35-45</td>
<td>&lt;1 %</td>
</tr>
<tr>
<td><strong>HYDROTHERMAL</strong></td>
<td></td>
<td></td>
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<tr>
<td>sericite</td>
<td>weakly disseminated pervasive</td>
<td>destroyed</td>
<td>plagioclase pseudomorphs</td>
<td>&gt;65</td>
<td>9 to 7 %</td>
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<tr>
<td>sericite+hematite+quartz</td>
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<td>montmorillonite-porphyrine</td>
<td>destruction by ser, locally CO3-chlorite pseudomorphs</td>
<td>&gt;70</td>
<td>20-60 %</td>
</tr>
<tr>
<td>quartz-sericite</td>
<td>pervasive with disseminated sulphides</td>
<td>destroyed</td>
<td>CO3-chlorite pseudomorphs</td>
<td>&gt;80</td>
<td>&lt;9 %</td>
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<tr>
<td>chlorite-sericite+pyrite+hematite</td>
<td></td>
<td></td>
<td>destroyed to moderately preserved</td>
<td>&gt;67</td>
<td>14 to 45 %</td>
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<tr>
<td>Mn-carbonate</td>
<td>spotty blebs with concentric and radiating internal structures</td>
<td>destroyed to moderately preserved</td>
<td>chl pseudomorphs</td>
<td></td>
<td></td>
</tr>
<tr>
<td>chlorite ± sphene</td>
<td>selective alteration of phenocrysts</td>
<td>destroyed to moderately preserved</td>
<td>chl pseudomorphs</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

of sills and the matrix of pumice breccias near Rosebery.

Some diagenetic alteration facies in the northern Central Volcanic Complex reflect the primary composition of the host volcanic facies, implying that diagenesis either involved fluid that was roughly in equilibrium with the rock or relatively low fluid/rock ratios. Hydrothermal alteration at Rosebery extends up to 100 m stratigraphically beneath the ore lenses, 10-20 m into the hangingwall and 500 m along strike in the footwall, commonly transgresses stratigraphic contacts and compositional boundaries and strongly modifying the primary composition (Large et al., in press b). Thus, hydrothermal activity at Rosebery probably involved interaction of fluid and rock with vastly different compositions and/or high fluid/rock ratios.

Mass gains and losses during diagenetic alteration of the northern Central Volcanic Complex were significantly smaller but more complicated than the changes related to footwall hydrothermal alteration at Rosebery. Total mass changes during diagenetic alteration were typically less than 5% but up to 9% (Fig. 6.8), whereas hydrothermal alteration at Rosebery varied from 5-50% (Fig 6.13, Appendix F). The different diagenetic alteration facies have been variably enriched or depleted in some trace elements over small distances, reflecting the complex pattern of overlapping assemblages on a small scale. In contrast, footwall hydrothermal alteration at Rosebery shows consistent enrichment in SiO2, K2O, Rb and depletions in CaO, Na2O, P2O5, Sr, Ba, Tl (Green et al., 1981; Naschwitz and Van Moort, 1991; Aung Pwa et al., 1992). Concentrations of MgO, MnO and Fe2O3 are dependent on the local proportions of the relevant alteration minerals (Green et al., 1981; Lees et al., 1990; Aung
Na$_2$O content and alteration indices also reflect the degree of plagioclase destruction and the nature of the alteration (Barrett et al., 1993). Unaltered volcanics typically have 2-5 wt % Na$_2$O (Barrett et al., 1993; Stolz et al., 1996b). Diagenetic alteration facies in rhyolitic units of the northern Central Volcanic Complex have 1-5 wt % Na$_2$O, although in strongly albitised samples Na$_2$O exceeds 5 wt %. Al and CCPI values range from 25-65 and 20-70 respectively. Hydrothermal alteration facies in the Rosebery footwall show Na$_2$O depletion (<2 wt % Na$_2$O, Large, 1992), high Al (>70) and moderate to high CCPI (40-80) values and plot in the dolomite-sericite-chlorite triangle of the Al-CCPI box plot (Fig. 6.7) (Large et al., in press a).

Figure 6.13: Mass changes during hydrothermal alteration associated with the Rosebery VHMS ore deposit. Mass changes were calculated using the method of Barrett and MacLean (1994) (Appendix F). A. and B. Absolute and relative mass changes during sericite and Mn-carbonate alteration zones. C. and D. Absolute and relative mass changes during quartz-sericite and chlorite alteration zones.

The products of diagenetic and hydrothermal alteration can have similar timing and mineralogy, and can significantly alter the appearance of glassy volcanic rocks. Specific criteria were used to discriminate alteration facies associated with diagenetic alteration in the northern Central Volcanic Complex and hydrothermal alteration associated with the VHMS mineralisation at Rosebery. These criteria included: distribution of the alteration facies, relationship with stratigraphic or textural boundaries, degree of textural modification and degree of chemical modification as reflected by mass change calculations, Na$_2$O content and alteration indices. These criteria could help discriminate diagenetic from hydrothermal alteration facies in other parts of the Mount Read Volcanics and other submarine volcanic successions.
6.7 Significance of recognising diagenetic alteration in a mineralised submarine volcanic succession

The recognition of diagenetic alteration and the discrimination between diagenetic and hydrothermal alteration in submarine volcanic successions have several implications for exploration and the study of ore genesis and alteration.

Textural changes indicate that large changes in porosity and permeability occurred as a result of cementation, alteration and compaction during diagenesis. These changes would have significantly influenced fluid pathways and the site of hydrothermal venting and mineralisation.

In many VHMS alteration studies, it has been assumed that diagenetic alteration has little or no effect on the whole-rock composition. However, this study suggests that compositional changes associated with diagenetic alteration can encompass mass changes up to 9%. Before changes in composition related to hydrothermal alteration can be assessed, changes associated with diagenetic alteration should be considered and eliminated.

Hydrothermal alteration zones surrounding VHMS deposits are a potential target for mineral exploration (Franklin et al., 1981). Thus the ability to discriminate between hydrothermal alteration facies related to mineralisation and regional diagenetic alteration facies is important for determining prospective from unprospective areas during exploration.

Timing relationships between diagenetic alteration facies and hydrothermal alteration facies at the margins of VHMS mineralisation can provide constraints on the timing and origin of mineralisation. This is particularly important for VHMS mineralisation that is interpreted to have formed by sub-seafloor replacement of glassy volcanic facies. In many cases, the processes of hydrothermal alteration and mineralisation are interpreted to be contemporaneous with diagenesis (Iijima, 1974, 1978; Utada, 1991; Allen, 1997) and overprinting alteration facies should reflect this.

Diagenetic mineral assemblages and the thickness of diagenetic alteration zones can also be used as an indicator of thermal history of the depositional basin (Utada, 1991). Depositional basins interpreted to have high geothermal gradients would be favourable for submarine hydrothermal mineralisation. In the northern Central Volcanic Complex, high geothermal gradients are reflected in high temperature diagenetic mineral assemblages, pervasive diagenetic alteration facies and thick, laterally extensive diagenetic alteration zones.

6.8 Conclusions

Present textures, mineral assemblages and compositions of the northern Central Volcanic Complex reflect the combined processes of diagenetic alteration, compaction, hydrothermal alteration, metamorphism and deformation. Metamorphosed diagenetic and hydrothermal alteration facies have been distinguished using a combination of alteration mineralogy, overprinting relationships, texture, distribution, intensity and whole-rock geochemistry.
The diagenetic alteration facies are characterised by: early formation, always preceding metamorphic assemblages and fabrics developed during regional deformation; regional distribution that is strongly influenced by the primary volcanic composition; complex distribution of alteration facies on a small scale (m's to cm's); generally good preservation or enhancement of primary textures, and feldspar crystals; relatively low alteration indices; and, small mass changes that were highly variable over short distances.

Diagnostic features of hydrothermal alteration in the northern Central Volcanic Complex are: local distribution; commonly pervasively developed, transgressing facies and unit contacts and primary textures; a lack of correlation between the hydrothermal alteration assemblage and the primary composition; a good correlation between the degree of textural modification and the alteration intensity; high alteration indices; substantial mass gains and losses of chemically mobile elements and consistent mass changes over wide areas (up to 500 m).

Diagenetic alteration facies in the northern Volcanic Complex record the formation of clay coatings, precipitation of pore-filling cements and the dissolution and replacement of volcanic glass synchronous with compaction. The earliest diagenetic alteration facies are sericite and carbonate in the felsic volcanic facies and chlorite-sericite-hematite in the mafic volcanic facies. These alteration facies probably replaced early clay minerals that coated glassy surfaces. Feldspar-quartz-sericite alteration (probably replacing diagenetic zeolites), filled primary pore space. Dissolution of glass was followed by chlorite-sericite and chlorite-sericite-hematite (after clay minerals and chlorite) replacement of originally glassy domains and feldspar-quartz-sericite alteration of both crystalline and glassy volcanic facies. This diagenetic sequence is similar to the alteration sequences in dominantly glassy volcaniclastic sands in modern back-arc basins (cf. Lee and Klein, 1986).

Post-compaction to syn-deformation chlorite-epidote alteration facies may reflect the direct conversion of glass and clays to chlorite and epidote during a stage transitional between diagenesis and metamorphism.

Hydrothermal alteration facies in the northern Central Volcanic Complex include: pre-compaction quartz-sericite and calcite-quartz-hematite alteration that was synchronous with diagenesis. Hydrothermal chlorite-pyrite, chlorite-calcite-magnetite and some quartz-sericite alteration are spatially associated with hydraulic breccias, deformational fabrics and structures.
Chapter 7
Synthesis: Cambrian palaeogeographic reconstruction and evolution of the northern Central Volcanic Complex

7.1 Introduction

In this chapter, the results of this study are summarised in a discussion of the Cambrian palaeogeographic setting and post-depositional evolution of the northern Central Volcanic Complex. Detailed analysis of volcanic (Chapters 3 and 5) and alteration facies (Chapter 6) in the northern Central Volcanic Complex, particularly east of the Mount Black Fault, has helped to constrain the facies architecture, depositional setting, stratigraphy and structure (Chapter 4). Significant advances have been made in understanding textural and compositional variations related to regional diagenetic alteration and in discriminating among diagenetic and hydrothermal alteration related to VHMS mineralisation (Chapter 6). Timing relationships between volcanism, alteration and deformation have been used to construct a model for the textural evolution of volcanic facies in the northern Central Volcanic Complex.

The results of this research are relevant to understanding other ancient submarine volcanic successions and the implications are discussed in this chapter. The similarities between the northern Central Volcanic Complex and other submarine volcanic successions hosting VHMS mineralisation are considered.

7.2 Palaeogeographic reconstruction of the northern Central Volcanic Complex

Here a palaeogeographic model for the volcanic and sedimentary history of the northern Central Volcanic Complex is discussed. This model describes the physical geography of the source and depositional areas during the Cambrian. It is reconstructed from regional facies analysis (Chapter 3) and interpretation of the stratigraphy and structure (Chapter 4). The palaeogeographic model comprises four stages: (1) growth of a submarine dominantly effusive, basaltic to dacitic volcano (the Sterling Valley Volcanics); (2) submarine effusion and intrusion of a large felsic volcanic complex (the Mount Black Formation); (3) a large felsic explosive eruption, probably accompanied by caldera subsidence (the Hercules and Kershaw Pumice Formations); and (4) emplacement of lavas and syn-volcanic intrusions (the Hercules and Kershaw Pumice Formations) broadly synchronous with the formation of massive sulfide mineralisation.

The salient points used to determine the character and setting of volcanic activity and to constrain the palaeogeographic model are revised (sections 7.2.1 to 7.2.4) before the model is presented (section 7.2.5).
7.2.1 Basement
The base of northern Central Volcanic Complex is not exposed. For the purpose of the palaeogeographic reconstruction, the northern Central Volcanic Complex is assumed to be the lowest part of the Mount Read Volcanics. The Mount Read Volcanics were deposited in an extensional basin floored by Late Proterozoic-Early Cambrian submarine sedimentary rocks and tholeiites (the Success Creek Group and the Crimson Creek Formation), and mafic-ultramafic complexes (Berry and Crawford, 1988; Crawford and Berry, 1992). Crystal-rich sandstone in the Mount Black Formation has a mixed provenance that includes the Precambrian basement, the Crimson Creek Formation and other volcanic sources. This implies that the Precambrian and Late Proterozoic-Early Cambrian rocks were exposed to erosion at the basin margins during the Middle Cambrian. The Success Creek Group and the Crimson Creek Formation are currently exposed to the west of the Rosebery Fault, which has been interpreted as a Cambrian, possibly syn-volcanic, growth fault (Corbett and Leces, 1987; Berry, 1991; Berry and Keele, 1997).

7.2.2 Sterling Valley Volcanics
The lowest stratigraphic unit in the northern Central Volcanic Complex is the Sterling Valley Volcanics. They are a thick (1.5 km) and laterally extensive (7 km) succession of intercalated dacitic to basaltic lavas, sills, dykes and volcaniclastic facies. This succession was emplaced in a submarine below-wave-base environment and derived from either a submarine or an emergent vent. The volcaniclastic facies were probably deposited by density modified grain flows, high-concentration turbidity currents and/or debris flows. Sparse water-settled laminated mudstone also occurs.

The Sterling Valley Volcanics are interpreted to represent the submarine apron of a large tholeiitic submarine volcano, probably located in an extensional basin that was developing synchronous with volcanic activity.

The Sterling Valley Volcanics are conformably overlain by the Mount Black Formation (formerly part of the Mount Black Volcanics). The gradational contact with the calc-alkaline Mount Black Formation and the presence in the Sterling Valley Volcanics of dacitic lavas and sills with calc-alkaline affinities suggests that coeval tholeiitic and calc-alkaline volcanism occurred, at least during deposition of the upper part of the Sterling Valley Volcanics.

7.2.3 Mount Black Formation
The Mount Black Formation (previously part of the Mount Black Volcanics) comprises a thick (>1.6 km) succession of rhyolitic and dacitic lavas, domes, cryptodomes and syn-volcanic sills. The volcaniclastic facies are mainly in situ, clast-rotated and reseledimented autobreccia and hyaloclastite. Rare water-settled mudstone and mass-flow-emplaced crystal-rich sandstone also occur.

The high-proportion of lavas and sills to volcaniclastic or sedimentary facies in the Mount Black Formation suggests proximity to source and that the Mount Black Formation constructed significant topographic relief, which influenced subsequent sedimentation patterns.

The occurrence of sparse laminated, pyritic black mudstone, abundant hyaloclastite, turbidites and other mass-flow-emplaced volcaniclastic facies suggest that eruption and deposition of the Mount...
Black Formation occurred below wave base. The construction of considerable topography implies that water depth varied considerably during emplacement of the Mount Black Formation.

The lateral extent of the Mount Black Formation is consistent with multiple vents, distributed over a 3 by 30 km area in the submarine basin, erupting compositionally similar magma. The Mount Black Formation is interpreted to represent a calc-alkaline submarine volcanic complex and its intrusive roots.

Pumice-breccia and pumice-lithic clast-rich breccia units of the Kershaw Pumice Formation conformably overlie the Mount Black Formation.

### 7.2.4 Kershaw and Hercules Pumice Formations

The Kershaw Pumice Formation (formerly part of the Mount Black Volcanics) is interpreted to be a correlate of the Hercules Pumice Formation (previously the Footwall Pyroclastics and Host Rocks), based on lithofacies characteristics, compositional similarities and stratigraphic position. These formations are dominated by non-welded rhyolitic pumice-rich facies intercalated with and intruded by rhyolitic and dacitic lavas and sills. The rhyolitic lavas and sills are commonly pumiceous at the margins and include pumiceous peperite and in situ and resedimented pumiceous hyaloclastite.

Angular, delicate pumice clasts and weakly graded, very thick (5-150 m) and laterally extensive beds of pumice breccia that lack other intercalated facies indicate that the entire pumice-rich formation was emplaced very rapidly, most likely synchronous with a major pyroclastic eruption. The bulk volume of pumiceous debris in the Kershaw Pumice Formation is estimated to be greater than 24 km$^3$, suggesting that eruption of the pumice-rich formations may have been accompanied by caldera subsidence (cf. Williams and McBirney, 1979; Lipman, 1984). Rhyolitic and dacitic sills compositionally similar to the pumice-rich facies intrude the Kershaw and Hercules Pumice Formations, reflecting the proximal, possibly intracaldera setting.

Lithofacies characteristics, the occurrence of locally intercalated black mudstone and associated VHMS mineralisation are interpreted to indicate deposition of the Kershaw and Hercules Pumice Formations in a submarine, below-wave-base environment. Although the precise water depth is poorly constrained, local occurrences of mudstone at the top of the Hercules and Kershaw Pumice Formations suggest that after deposition, the top of the formations remained below wave base. Thus, taking the thickness of the Hercules and Kershaw Pumice Formations (>800 m) into account, the implication is that eruption occurred in deep water (at least 1000 m) or that syn-eruptive subsidence occurred. Theoretical constraints for depths at which large scale felsic eruptions can occur (McBirney, 1963) and descriptions of modern submarine pyroclastic eruptions and calderas (Fiske et al., in press) suggest that such eruptions may have occurred up to 900 m below sea level but are unlikely to have occurred at significantly greater depths. The volume of the pumice-rich facies association in the Kershaw and Hercules Pumice Formations and syn-depositional faults in the Hercules Pumice Formation (Allen, 1993 unpub., 1994a unpub.) are consistent with a caldera-forming eruption. Although an intracaldera setting for the Kershaw and Hercules Pumice Formations is highly plausible, the original margins of the caldera have not been located.
Stratified pumice-rich sandstone and shard-rich siltstone (Host-rock member) at the top of the pumice-rich association are interpreted to represent the last stages of syn-eruptive sedimentation from the pyroclastic eruption, as pumice, shards and crystals were deposited either by suspension settling of pyroclasts that have accumulated in the water during eruption or from turbidity currents. There is a gradational change from interbedded syn-eruptive pumice-rich facies to slightly reworked sandstone and siltstone that also contain rare quartz crystals from an extrabasinal source (Allen, 1994a unpub.).

The last stage in the evolution of this volcanic centre was the emplacement of rhyolitic and dacitic lavas, domes and syn-volcanic sills. Lavas and sills commonly have pumiceous margins and associated pumiceous autochthonous facies. Laterally discontinuous pumice-lithic clast-rich breccia at the top of the Kershaw Pumice Formation may reflect redistribution of pumice clasts and non-vesicular lithic clasts derived either from rhyolitic domes or from syn-volcanic faults scarps.

Carbonate facies are common towards the top of the Kershaw Pumice Formation and are interpreted to be the product of Cambrian sub-seafloor hydrothermal alteration of volcaniclastic facies.

Quartz-phyric volcaniclastic facies and pyritic black mudstone of the White Spur Formation overlie the Hercules Pumice Formation. This contact is locally conformable and disconformable (Corbett and Solomon, 1989; R.L. Allen written communication, 2000).

### 7.2.5 White Spur Formation

The White Spur Formation is a 5-400 m-thick succession of massive to stratified feldspar-quartz-phyric, crystal-rich and pumice-rich volcaniclastic units interbedded with sandstone and black pyritic mudstone (Corbett and Lees, 1987; Lees, 1987; Allen, 1991 unpub.; McPhie and Allen, 1992). These facies are interpreted to be deposited in a below-wave-base environment from an adjacent rhyolitic volcanic centre (Allen, 1991 unpub.). They indicate a significant change in provenance, which may reflect a considerable time gap at the end of the Central Volcanic Complex.

### 7.2.6 Formation of the northern Central Volcanic Complex

Four stages are interpreted in the formation of the northern Central Volcanic Complex.

STAGE 1 (Fig. 7.1): The Sterling Valley Volcanics, a large tholeiitic basaltic submarine volcano, are formed in an extensional basin. They mainly consist of basaltic lavas, sills and dykes, and associated resedimented volcaniclastic debris. Volcaniclastic facies were deposited by density modified grain flows, high-concentration turbidity currents and/or debris flows. Late in Stage 1 calc-alkaline dacite lavas were erupted and interfinger with the products of the basaltic volcano. At the margins of the volcanic centre, black mudstone and volcanic siltstone were being deposited from suspension in the water column and low-concentration turbidity currents.

STAGE 2 (Fig. 7.2): Eruptions from seafloor intrabasinal vents build a broad volcanic edifice consisting of overlapping felsic volcanoes (Mount Black Formation). These volcanoes mainly comprise rhyolitic and dacitic lavas, domes, cryptodomes and syn-volcanic sills. Some autobreccia and hyaloclastite
LEGEND

- Resedimented felsic autobreccia and hyaloclastite
- Dacitic lavas, domes, cryptodomes and syn-volcanic sills
- Basaltic lava
- Resedimented mafic hyaloclastite, pillow breccia fragments and scoria
- Sterling Valley Volcanics
- Crimson Creek Formation
- Success Creek Group
- Precambrian basement
- Quartz-phyric turbidites
- Quartz-feldspar-phyric rhyolitic syn-volcanic sills
- Resedimented felsic autobreccia, hyaloclastite and pumice clasts
- Shard-rich sillstone
- Non-welded pumice breccia
- Mixed provenance crystal-rich turbidites
- Rhyolitic lavas, domes, cryptodomes and syn-volcanic sills
- Massive sulfide mineralisation

Figure 7.1: Schematic palaeogeographic reconstruction and facies model for the tectonic, volcanic and sedimentary setting of the northern Central Volcanic Complex and source terrain during eruption of the Sterling Valley Volcanics. 1 - construction of a large submarine volcano, composed of basaltic lavas and volcaniclastic facies; 2 - dacitic lava and associated resedimented hyaloclastite; 3 - dacitic domes.

Associated with these facies is resedimented by down slope rolling, dome collapse, grain flow and debris flow. Crystal-rich sandstone derived from a more distal quartz-phyric source, exposed Precambrian basement and Crimson Creek Formation is deposited in the basin at the flanks of the growing volcanic complex.

STAGE 3 (Fig. 7.3): Eruption of rhyolitic pumice-rich facies association of the Kershaw and Hercules Pumice Formations synchronous with or immediately prior to caldera subsidence. The water column is choked with glassy ash and temporarily buoyant pumice. Deposition in the proximity of the vent of water-logged pumice clasts, shards and crystals is by water-supported mass flows and water-settled fall.
Figure 7.2: Schematic palaeogeographic reconstruction and facies model for the tectonic, volcanic and sedimentary setting of the northern Central Volcanic Complex and source terrain during emplacement of the Mount Black Formation. This involved effusive eruptions and the intrusion of a large felsic volcanic complex. 1 - dacitic and rhyolitic lavas and domes; 2- dacitic syn-volcanic intrusion; 3 - resedimented autobreccia or hyaloclastite associated with the growth of a rhyolitic dome; 4 - syn-eruptive deposition of volcaniclastic breccia derived from dome collapse; 5 - post-eruptive resedimentation of extrabasinally-derived crystal-rich sandstone. See Figure 7.1 for legend.

STAGE 4 (Fig. 7.4): Post-caldera, rhyolitic and dacitic cryptodomes and syn-volcanic sills of the Kershaw and Hercules Pumice Formations intrude the thick, probably intracaldera, pumice breccia. Locally rhyolites breach the surface forming lavas and domes with thick carapaces of in situ, clast-rotated and resedimented pumiceous hyaloclastite. Slumping from syn-volcanic fault scarps and resedimentation from active rhyolitic lavas and domes forms pumice- lithic clast-rich breccia and stratified sandstone on the caldera floor. Volcaniclastic turbidites derived from an extrabasinal rhyolitic volcanic centre begin to deposit quartz-phryic volcaniclastic facies in the Central Volcanic Complex basin. This stage is marked by the onset of significant hydrothermal activity. Hydrothermal fluid infiltrates and replaces pumice-rich facies sub-seafloor and is exhaled onto the seafloor depositing massive sulfides.

7.3 Post-eruptive evolution of the northern Central Volcanic Complex

Current mineral assemblages and textures in the northern Central Volcanic Complex reflect the processes of diagenesis, hydrothermal alteration, compaction, metamorphism and deformation.
Figure 7.3: Schematic palaeogeographic reconstruction and facies model for the tectonic, volcanic and sedimentary setting of the northern Central Volcanic Complex and source terrain during deposition of the Kershaw and Hercules Pumice Formation. Explosive eruption was synchronous with or immediately followed by caldera subsidence. 1 - syn-eruptive pumice breccia; 2 - caldera subsidence; 3 - syn-eruptive sedimentation of shards and pumice clasts by water-settling. See Figure 7.1 for legend.

Discriminating among the products of diagenetic and hydrothermal alteration and metamorphism in the northern Central Volcanic Complex was achieved using detailed analysis of the alteration textures, distributions and overprinting relationships at a number of different scales (Chapter 6). The results have helped to constrain the textural and compositional history of the northern Central Volcanic Complex. Diagenetic alteration began shortly after eruption (probably within 1 Ma) in the Cambrian and continued until the transition to regional metamorphism. Prior to burial compaction, but synchronous with diagenesis and the intrusion of syn-volcanic sills, hydrothermal alteration associated with VHMS mineralisation at Rosebery and Hercules commenced (Allen, 1997 unpub.). The transition from diagenetic alteration to regional metamorphism is poorly defined and is assumed to be gradational with increasing temperature and pressure as a result of burial. Peak regional metamorphism was synchronous with Devonian deformation and is overprinted by contact metamorphic aureoles associated with the emplacement of Devonian granites.

These post-eruptive changes in the northern Central Volcanic Complex are synthesised in a genetic model. This model comprises six successive stages in the evolution of the northern Central Volcanic Complex (Fig. 7.5): (1) the onset of diagenesis; (2) emplacement of syn-volcanic sills, hydrothermal alteration and mineralisation; (3) dissolution and alteration of glass synchronous with compaction; (4)
Figure 7.4: Schematic palaeogeographic reconstruction and facies model for the tectonic, volcanic and sedimentary setting of the northern Central Volcanic Complex and source terrain during the final stages of deposition of the Kershaw and Hercules Pumice Formation and formation of massive sulfide mineralisation. 1 - syn-volcanic rhyolitic and dacitic sills; 2 - syn-volcanic cryptodome resulting in up-doming and resedimentation of the overlying sediments; 3 - local redeposition of the top of the water-settled sediments by turbidity currents; 4 - emplacement of rhyolitic lavas and associated syn-eruptive resedimented autoclastic facies including pumiceous hyaloclastite; 5 - resedimentation of debris from fault scarps; 6 - influx of quartz-phyric turbidites from a distal rhyolitic volcanic centre; 7 - sub-seafloor infiltration and replacement by massive sulfide mineralisation; 8 - exhalation of hydrothermal fluids onto the seafloor and construction of a massive sulfide mound. See Figure 7.1 for legend.

replacement of early authigenic minerals and remaining glass by more stable mineral assemblages; (5) Devonian regional metamorphism and deformation; and (6) emplacement of granites and associated metamorphism and alteration.

7.3.1 STAGE 1: The onset of diagenesis
Stage 1 summarises the onset of diagenetic alteration and includes both the coating of original surfaces, the infilling of primary porosity and replacement of glass by new minerals. The earliest alteration products in the northern Central Volcanic Complex are thin films of sericite, chlorite-sericite-hematite and carbonate which coat all original surfaces, including vesicle walls, plagioclase crystals, shards and fractures (Fig. 7.6). These coatings are interpreted to be the metamorphic equivalent of smectite clays, carbonate and palagonite, which mark the beginning of diagenetic alteration.
STAGE 1: The onset of regional, syn-volcanic diagenesis. 1 - Original surfaces were coated in new minerals and zeolites and K-feldspar filled primary pore space. 2 - Fluid derived from interstitial pore water and seawater trapped in the volcanic succession migrated towards the seafloor.

STAGE 2: Hydrothermal alteration synchronous with continuing regional diagenetic alteration. 1 - Syn-volcanic sills intruded unconsolidated sediments causing local alteration. 2 - Hydrothermal alteration associated with mineralisation. Hydrothermal fluid flow was unfocused and ponded beneath sills (Allen, 1994b unpub.). 3 - Massive sulfide mineralisation formed by sub-seafloor replacement and exhalation onto the seafloor.

STAGE 3: Diagenetic alteration and compaction, synchronous with deposition of the White Spur Formation (1). 2 - Clays, chlorite and sericite replaced originally glassy domains and were flattened during compaction (fiamme). At depth zeolites, plagioclase and K-feldspar were replaced by albite. 3 - Seawater migrated laterally along porous units. 4 - Hangingwall hydrothermal alteration occurred in the White Spur Formation.

Figure 7.5: Model for the post-depositional evolution of the northern Central Volcanic Complex. Schematic cross sections through the northern Central Volcanic Complex are not to scale. WSF = White Spur Formation.
STAGE 4: Transition between diagenesis and metamorphism with increasing depth of burial. 1 - Remaining glassy domains and authigenic minerals were replaced by more stable minerals. 2 - Migration of fluid released from dehydration reactions of clay minerals. Chlorite-epidote alteration zone developed at depth.

STAGE 5: Regional metamorphism and deformation (Devonian). Mineral assemblages and textures were overprinted by greenschist facies mineral assemblages and tectonic fabrics. Regional folding, faulting and shearing, including reconfiguration of the massive sulfide lenses. Restricted, structurally-controlled fluid flow.

STAGE 6: Emplacement of a shallow-level Devonian granite and associated metasomatic alteration. Fluid flow was strongly controlled by regional structures, particularly faults and shear zones.

Figure 7.5 continued: Model for the post-depositional evolution of the northern Central Volcanic Complex. Not to scale.

The formation of smectites and palagonite occurred at low temperatures between 0 and 120°C (Iijima and Utada, 1971; Iijima, 1978; Lee and Klein, 1986; Utada, 1991). Above these temperatures smectite would have been converted to sericite and illite (Simmons and Browne, 1996).

Post-dating the development of clay films, and pre-dating the compaction foliation (S1), the precipitation of cements filled original pore space, particularly vesicles and perlitic fractures (Fig. 7.6). This stage is characterised by feldspar-quartz-sericite, chlorite-sericite and chlorite-sericite-hematite alteration facies. Locally preserved relic textures in the feldspar-quartz-sericite alteration facies suggest that fibrous radiating zeolites originally filled vesicles and replaced the glassy kernels in perlite. Later in the diagenetic history, K-feldspar and albite extensively replaced zeolites. This stage also involved the nucleation of K-feldspar overgrowths on plagioclase phenocrysts (Fig. 7.6).
STAGE 1: thin films (in gold) of smectite clay, carbonate, palagonite or Fe-oxide coat all original surfaces, including vesicle walls, plagioclase crystals and shards.

STAGES 1 & 2: the primary porosity was filled and originally glassy shards and vesicle walls were replaced by zeolites (cream) and K-feldspar (orange). K-feldspar alteration nucleated on plagioclase phenocrysts and spread outwards.

STAGES 3 & 4: Stage 3 comprised the dissolution and alteration of glass synchronous with compaction. Fiamme were produced by chlorite-sericite and clay (green) alteration of originally glassy domains and mechanical compaction. Clays and Fe-oxides precipitated synchronous with the dissolution of glass, forming a stylolitic foliation (black). In Stage 4, any remaining glass and pre-existing clays and zeolites were replaced by more stable minerals (chlorite, sericite, albite and epidote). Zeolites, K-feldspar and plagioclase were extensively replaced by albite (yellow).

STAGE 5: Regional metamorphism and deformation modified primary volcanic and diagenetic alteration assemblages and textures. In areas of intense deformation, the diagenetic compaction foliation ($S_1$ fiamme and stylolites) was crenulated and stretched into the regional cleavage ($S_2$).

Figure 7.6: Schematic model for the microscopic textural evolution and reduction of porosity in pumice-rich facies in the northern Central Volcanic Complex during four stages of diagenesis and regional metamorphism and deformation.
The infilling of primary pore space had a dramatic effect on the porosity of pumice-rich facies. Henneberger and Browne (1988) found that during zeolite facies alteration (mordenite-clinoptilolite alteration) the porosity of pumice-rich facies was reduced by half (from 34–47 \% to 20–50 \%). Alteration in the quartz-adularia zone further reduced porosity (to 4–23 \%). Pumice clasts in the Kershaw and Hercules Pumice Formations are mainly interpreted to be the products of pyroclastic eruptions and are inferred to have had primary porosities between 60–90 \% (cf. Sparks, 1978; Houghton and Wilson, 1989; Thomas et al., 1994). Unlithified, non-welded pumice-rich breccias, similar to the pumice breccias in the northern Central Volcanic Complex, can be estimated to have a mean porosity of 50–70 \% (cf. Allen, 1987). In contrast, the porosity of coherent facies would have initially been less than 10\% and possibly as low as 1\% (cf. Davis, 1966).


These early stages of diagenesis probably involved fluid derived from interstitial pore water and seawater trapped in the volcanic succession (cf. Boles and Coombs, 1977). This modified seawater would have been expelled from the volcanic succession in response to overburden pressure and migrated towards the seafloor as diffuse unfocused flow.

7.3.2 STAGE 2: Emplacement of syn-volcanic sills, hydrothermal alteration and mineralisation synchronous with diagenesis

Two types of early hydrothermal alteration are recognised in the northern Central Volcanic Complex. The first is associated with the emplacement of syn-volcanic sills, and the second is associated with VHMS mineralisation.

Emplacement of syn-volcanic sills: Syn-volcanic sills emplaced into the northern Central Volcanic Complex typically have peperitic contacts consistent with the intrusion of magma into unconsolidated sediment. The host facies within several tens of metres of these sills has been indurated and quartz-sericite-altered (Fig. 7.5), significantly reducing the porosity adjacent to the sill. In addition, thermal pulses brought about by the intrusion of sills may have influenced fluid circulation patterns and raised the local geothermal gradient (cf. Einsele et al., 1980; Utada, 1991).

Mineralisation and hydrothermal alteration: The Rosebery massive sulfide deposit comprises stacked ore lenses with associated alteration haloes. The quartz-sericite, Mn-carbonate, chlorite and sericite hydrothermal alteration haloes pre-date compaction and were synchronous with diagenesis (Large et al., in press b).

Massive sulfide mineralisation at Rosebery and Hercules is hosted in thick, mass-flow-emplaced pumice breccia and sandstone units near the top of Hercules Pumice Formation. This implies that mineralisation formed by replacement and impregnation of originally porous volcaniclastic facies (Allen, 1990b unpub.; Khin Zaw and Large, 1992; Allen 1994b unpub). The nature of the host pumice-rich facies, the volcanic architecture through which hydrothermal fluids ascended, and the distribution
of pre-existing alteration facies influenced the distribution and geometry of mineralisation and their associated alteration haloes.

Syn-volcanic sills above and below the ore lenses intruded both prior to and possibly after mineralisation (Allen, 1994b unpub.). These sills commonly have peperitic margins which implies that the host facies was unconsolidated and possibly wet at the time of intrusion (Chapter 5 and White et al., 2000). The originally highly porous, water-saturated and glassy nature of the pumice breccia in the Hercules Pumice Formation allowed hydrothermal fluids to be poorly focused. Ascending hydrothermal fluids would have permeated through the substrate producing widespread stratabound alteration zones and sheet-like lenses of mineralisation (Large, 1992). Sulfide deposition may have occurred at the interface between ascending hydrothermal fluids and overlying cold, seawater-saturated strata (Allen, 1994b unpub.; Khin Zaw and Large, 1992).

The occurrence of syn-volcanic sills near the top of the Hercules Pumice Formation prior to mineralisation would have provided a relatively impermeable barrier, comprising indurated host facies and/or solidified sill. This may have caused ponding of fluid beneath the sill and promoted sub-seafloor mineralisation by replacement and infiltration of the sill margin and host sediments (Fig. 7.5) (Allen, 1994b unpub.).

The distance below the seafloor at which impregnation occurred is poorly constrained. The occurrence of massive sulfide clasts in mass-flow units in the base of the White Spur Formation (Lees, 1987; Corbett and Lees, 1987; Corbett and Solomon, 1989; Corbett, 1992; McPhie and Allen, 1992) suggests that sulfide deposition probably occurred within several metres of the seafloor and/or that hydrothermal fluids locally exhaled onto the seafloor (Allen 1994b unpub.). The occurrence of pumiceous peperite at the margins of many of the sills interpreted to be approximately synchronous with mineralisation may help constrain the water depth and thickness of the sediment cover during mineralisation. Pumiceous peperite probably formed at shallow depths in the subsurface at less than 200 m below sea level (Chapter 5) (McBirney, 1963; Hunns and McPhie, 1999).

The formation of massive sulfide deposits at Rosebery and Hercules required a large active hydrothermal system that was probably sustained by a high geothermal gradient. In the northern Central Volcanic Complex, a high geothermal gradient is supported by: the distribution, thickness and mineralogy of the diagenetic alteration zones, proximity to volcanic vents, and the interpreted setting within an extensional tectonic regime. In the northern Volcanic Complex, there are two regionally developed, thick, originally flat-lying, diagenetic alteration zones: an albite zone and an epidote zone. Mineral assemblages in these zones reflect the reaction of glass with interstitial porewater at a particular temperature. The albite zone is characterised by feldspar-quartz-sericite and chlorite-sericite alteration facies and probably formed at temperature between 100-190°C (cf. Iijima and Urada, 1971; Thompson, 1971; Merino, 1975; Munha et al., 1980; Boles, 1982; Lee and Klein, 1986). The epidote zone is characterised by chlorite-epidote, chlorite-sericite-hematite and feldspar-quartz-sericite and formed at temperatures of at least 200°C (cf. Seki, 1972; Kristmannsdottir, 1979). The thick (2 km), high-temperature albite alteration zone is consistent with a high-grade diagenetic alteration system and elevated geothermal gradient. The magma chamber that supplied the Mount Black Formation and the Kershaw and Hercules Pumice Formations was large and would have provided significant heat, elevating
the geothermal gradient and promoting circulation of hydrothermal fluid.

Mineralisation and associated hydrothermal alteration haloes at Rosebery and Hercules are interpreted to have formed at temperatures greater than 300°C (Gemmell and Large, 1992; Large, 1992). Ongoing regional diagenetic alteration was probably at temperatures up to 150°C.

### 7.3.3 STAGE 3: Diagenetic alteration and compaction

Continuing diagenetic alteration involved the dissolution and alteration of glass synchronous with compaction. This stage is characterised by chlorite-sericite and chlorite-sericite-hematite assemblages replacing originally glassy shards, clasts and domains. The occurrence of the spaced chlorite-sericite-hematite stylolitic dissolution fabric suggests that glass was dissolved and clays and Fe-oxides precipitated synchronous with compaction (Fig. 7.6). Subsequently clays were replaced by sericite and chlorite. Albite dissolution of plagioclase phenocrysts and replacement of earlier (Stage 1) zeolites and K-feldspar by albite were probably synchronous with phyllosilicate alteration and compaction during this stage.

Widespread albite dissolution associated with diagenesis in modern geothermal systems occurs at measured temperatures of between 100-190°C (Hiijima and Utada, 1971; Thompson, 1971; Merino, 1975; Munha et al., 1980; Boles, 1982).

In pumice breccias in the northern Central Volcanic Complex, the initially high porosity (50-70%) was reduced prior to compaction by precipitation of clays and zeolites in the primary pore spaces. This would have strengthened the grain framework and reduced the subsequent mechanical compaction.

Compositional changes associated albite and phyllosilicate alteration resulted from the interaction of felsic glass with seawater (cf. Boles and Coombs, 1977; Hajash and Chandler, 1981; Boles, 1982; Shiaki and Iiyama, 1990). Large volumes (>100 cm$^3$/cm$^2$) of pore water were probably displaced as a result of compaction under the weight of accumulating sediments (cf. Bethke, 1989; Einsele, 1992). Rapid and variable sedimentation rates during deposition of the White Spur Formation may have overpressured the pore fluid, promoting lateral fluid flow along permeable layers (cf. Galloway, 1984; Bethke, 1989).

Synchronous with diagenetic alteration and compaction in the Central Volcanic Complex was the onset of diagenetic alteration and hydrothermal alteration in the White Spur Formation. Weak hangingwall alteration developed during continued hydrothermal activity (cf. Sato et al., 1979).

### 7.3.4 STAGE 4: Transition between diagenesis and metamorphism

Early alteration minerals and remaining glassy domains were replaced by more stable mineral assemblages. This is interpreted to be a transitional stage between the end of diagenesis and compaction, and the onset of regional metamorphism. It is characterised by chlorite-epidote alteration facies in mafic and felsic volcanic facies in the deepest part of the northern Central Volcanic Complex (Fig. 7.5).

The isochemical nature of this alteration facies (Chapter 6) and stratigraphic control suggest the alteration was in response to increasing geothermal gradient with increasing depth of burial. This is
consistent with the high (>200°C) temperature epidote-rich assemblage (Seki, 1972; Kristmannsdottir, 1979). In these conditions, water is commonly released by dehydration of clay minerals or other hydrous minerals (Galloway, 1984).

7.3.5 STAGE 5: Regional metamorphism and deformation
Overprinting the diagenetic and hydrothermal alteration mineral assemblages and textures are greenschist facies mineral assemblages and tectonic fabrics associated with the Devonian metamorphism and deformation. During metamorphism, clays and any remaining zeolites were replaced by greenschist facies mineral assemblages including sericite, epidote, calcite and actinolite. Regional metamorphism was largely isochemical involving little or no compositional changes.

Devonian deformation modified primary volcanic and alteration textures and resulted in folds, faults and shear zones in northern Central Volcanic Complex. It also caused significant changes in the overall configuration of the originally stratabound ores lenses and underlying stringer zone at Hercules and the southern end of Rosebery (Fig. 7.5) (Khin Zaw, 1991).

Assemblages of calcite, epidote, tremolite-actinolite in some of the intermediate and mafic rocks in the Mount Read Volcanics imply that the regional temperature gradient was 370-450°C during metamorphism (Green et al., 1981).

The distribution of syn-S2 chlorite-pyrite and chlorite-calcite-magnetite alteration facies (Chapter 6) suggests that metamorphic fluid migration was restricted and strongly controlled by regional structures, particularly faults and shear zones.

7.3.6 STAGE 6: Emplacement of Devonian granites and hydrothermal alteration
Post-dating regional metamorphism and deformation, Devonian granite intruded the Rosebery mine area. The presence of granite at depth is indicated by gravity data (Leaman and Richardson, 1989), the recrystallisation and textural modification of ore minerals, and the growth of new minerals in ore lenses at the south end of Rosebery mine (Khin Zaw, 1991). New minerals formed in the contact aureole include pyrrhotite, pyrite, garnet, helvite, biotite and tourmaline (Solomon et al., 1987). Isotopic studies suggest that temperatures associated with emplacement of granite exceeded 370°C (Corbett and Solomon, 1989; Khin Zaw, 1991).

Abundant quartz ± calcite ± tourmaline ± magnetite veins interpreted to be associated with intrusion of Devonian granites suggest that fluid flow was structurally controlled.

7.4 Analogues for the northern Central Volcanic Complex
Volcanic facies, processes and volcano types represented in the northern Central Volcanic Complex have analogues in younger marine volcanic successions that host massive sulfide mineralisation. The northern Central Volcanic Complex records styles of volcanism and mineralisation similar to those in the Manus Basin and the Hokuroku Basin (Table 7.1). The Central Volcanic Complex, Manus Basin and Hokuroku Basin are interpreted to have formed in back-arc extensional settings (Chapter 4) (Ohmoto, 1983; Ohmoto and Skinner, 1983; Gamo et al., 1997; Waters and Binns, 1998; Yeats and
Table 7.1: Comparison of the main features of the northern Central Volcanic Complex, the Manus Basin and the Hokuroku Basin.

<table>
<thead>
<tr>
<th>Location</th>
<th>northern Central Volcanic Complex</th>
<th>Manus Basin</th>
<th>Hokuroku Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>Mount Read Volcanics</td>
<td>Papua New Guinea</td>
<td>Green Tuff Belt, Japan</td>
</tr>
<tr>
<td>Age</td>
<td>Cambrian (~500 Ma)</td>
<td>active</td>
<td>Miocene (~11-16 Ma)</td>
</tr>
<tr>
<td>Tectonic setting</td>
<td>extensional, back-arc basin</td>
<td>extensional, back-arc basin</td>
<td>extensional, back-arc basin</td>
</tr>
<tr>
<td>Basement</td>
<td>thinned passive margin, ocean floor, basalts and sediments</td>
<td>thinned basaltic island arc crust</td>
<td>thinned, extensional, back-arc basin</td>
</tr>
<tr>
<td>Environment of deposition</td>
<td>submarine, below-wave-base</td>
<td>submarine, below-wave-base (1200-2000 m)</td>
<td>submarine, below-wave-base (2000-3000 m)</td>
</tr>
<tr>
<td>Composition of volcanic succession</td>
<td>basaltic to rhyolitic with calc-alkaline and tholeiitic affinities</td>
<td>proximal lavas, in situ, clast-rotated and reseedimented autobreccia and hyaloclastite, pumice breccias, pillow lavas and scoria breccias</td>
<td>consistent with a fractionation series, coherent and autoclastic facies of lavas, domes, cryptodomes and syn-volcanic sills, pumice-rich facies, pumice-lithic clast-rich breccia and sparse mudstone and extrabasinal sandstone</td>
</tr>
<tr>
<td>Principal facies</td>
<td>coherent and autoclastic facies of lavas, domes, cryptodomes and syn-volcanic sills, pumice-rich facies, pumice-lithic clast-rich breccia and sparse mudstone and extrabasinal sandstone</td>
<td>explosive eruptions, and large explosive eruption/s</td>
<td>effusive, submarine caldera eruptions, explosive eruptions from emergent vents</td>
</tr>
<tr>
<td>Styles of volcanism</td>
<td>mainly effusive eruptions and shallow intrusions and large explosive eruption/s</td>
<td>hydrothermal alteration</td>
<td>diagenetic and hydrothermal alteration</td>
</tr>
<tr>
<td>Post-depositional changes</td>
<td>diagenetic and hydrothermal alteration, metamorphism and deformation</td>
<td>diagenetic and hydrothermal alteration</td>
<td>vertical zonation from glast to albite zone</td>
</tr>
<tr>
<td>Diagenetic grade</td>
<td>thick, high-grade diagenetic alteration zones (albite-epidote)</td>
<td>seafloor alteration, manganese coatings</td>
<td>alteration envelopes</td>
</tr>
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<td>Metamorphic grade</td>
<td>greenschist facies</td>
<td>Cu-Pb-Zn-Au-Ag</td>
<td></td>
</tr>
<tr>
<td>Hydrothermal alteration</td>
<td>roughly stratiform alteration halos</td>
<td>Pb-Zn-Cu</td>
<td>Cu-Pb-Zn-Au-Ag</td>
</tr>
<tr>
<td>Metals</td>
<td>Pb-Zn-Cu</td>
<td>sphalerite+galena+chalcopyrite+pyrite</td>
<td>galena+sphalerite+barite+chalcopyrite+pyrite</td>
</tr>
<tr>
<td>Main ore minerals</td>
<td>sphalerite+galena+chalcopyrite e+pyrite</td>
<td>sphalerite+galena+chalcopyrite+pyrite</td>
<td>massive sulfide mounds, chimneys, disseminated pyrite</td>
</tr>
<tr>
<td>Style of deposits</td>
<td>stratiform sheet-like lenses</td>
<td>massive sulfide mounds, chimneys, disseminated pyrite</td>
<td>mounds and conformable massive sulfide lenses and underlying utnger or disseminated ore</td>
</tr>
<tr>
<td>Origin of mineralisation</td>
<td>sub-seafloor replacement and exhalation 5,6,7</td>
<td>exhalative, minor replacement</td>
<td>exhalative, disseminated replacement</td>
</tr>
<tr>
<td>Host facies</td>
<td>pumice breccia</td>
<td>dacitic coherent and autoclastic facies</td>
<td>laves, pumice-rich facies and mudstone</td>
</tr>
</tbody>
</table>

All three comprise variable proportions of rhyolitic to basaltic, coherent and volcaniclastic facies deposited in a below-wave-base submarine environment (Urabe, 1979 unpub.; Guber and Merrill, 1983; Tanimura et al., 1983; Nakajima, 1988; Binns and Scott, 1993; Gamo et al., 1997; Waters and Binns, 1998; Yeats and Binns, 1998). In addition, they are characterised by an association of massive sulfide mineralisation with felsic volcanism (Horikoshi, 1969; Binns and Scott, 1993).

Comparisons between the Central Volcanic Complex, Manus Basin and Hokuroku Basin have several limitations because of post-depositional changes in the northern Central Volcanic Complex. Modification by folding, faulting and erosion means that understanding of the volcanic facies and facies architecture is incomplete and that the extent and distribution of the northern Central Volcanic Complex is poorly constrained.

7.4.1 Manus Basin, PNG

The Manus Basin, Papua New Guinea, is located NW of the active New Britain volcanic arc (Yeats and Binns, 1998). The Manus Basin is a large (>500 km long) submarine basin which includes a number of NE-trending basaltic to dacitic volcanic edifices (Fig. 7.7) (Yeats and Binns, 1998; Auzende et al., 2000). Typically these volcanoes rise 1200 m above the seafloor at 2200 m (Waters and Binns, 1998; Auzende et al., 2000). They are mainly composed of lavas comprising coherent facies, in situ, clast-rotated and resedimented autobreccia and hyaloclastite (Binns and Scott, 1993; Waters and Binns,
Pumice clasts collected from the seafloor have been interpreted to be pumiceous autobrecia and hyaloclastite (Waters and Binns, 1998). Felsic pyroclastic eruptions have been associated with the emergence of a large rhyolitic seafloor volcano (Tuluman volcano, Reynolds and Best, 1976) and the formation of calderas at Rabaul and Tavui in the adjacent New Britain volcanic arc (Tiffin et al., 1990). Several submarine calderas, with diameters up to 40 km, have also been recognised on the summits of isolated mafic volcanoes in the western Manus Basin (Gamo et al., 1997; Auzende et al., 2000).

The Manus Basin includes four main hydrothermal areas where massive sulfide is actively forming on the seafloor: Marmin Knolls, PACMANUS, DESMOS and Susu Knolls, (Yeats and Binns, 1998). Hydrothermal activity associated with massive sulfide mineralisation occurs at depths between 1200 and 1930 m and is hosted by mafic (Marmin Knolls), intermediate (DESMOS) and felsic (Susu Knolls and PACMANUS) volcanic facies (Binns and Scott, 1993; Gamo et al., 1997; Yeats and Binns, 1998; Auzende et al., 2000).

The Manus Basin, particularly the Paul Ridge (PACMANUS) and Susu Knolls areas are proposed as a modern analogue for the parts of the Central Volcanic Complex, mainly the Mount Black Formation and the Sterling Valley Volcanics. However, these comparisons have limitations as volcanic facies analysis in the Manus Basin is very limited, with only a small part of the basin being actively explored. In addition, research has relied on dredge samples and submersible dives, thus restricting understanding to the surface geology.

Similarities between the northern Central Volcanic Complex and the Manus Basin include:

(1) both successions comprise dominantly coherent and autoclastic facies with a broad range in composition from basalt to rhyolite or rhyodacite but with a predominance of dacite;

(2) the common occurrence of pumiceous autoclastic facies at the margins of lavas;

(3) they contain a lower successions of submarine basaltic and andesitic lavas and resedimented autoclastic facies overlain by dacitic lavas and domes.

Notable differences are:

(1) The Manus Basin is an order of magnitude larger than the preserved size of the northern Central Volcanic Complex (30 x 7 km);

(2) In the Manus Basin basalts and dacites appear to be related by a fractionation series (Binns and Scott, 1993). In the northern Central Volcanic Complex dacites have calc-alkaline affinities, whereas basalts are tholeiitic, implying that they were derived from different parent magmas (Chapter 4).

(3) The northern Central Volcanic Complex contains thick pumice breccias derived from a large explosive eruption. Although, large felsic submarine caldera structures have been recognised in the Manus Basin and submarine calderas and their associated pumice-rich deposits are present in the New
Britain volcanic arc (Tiffin et al., 1990 unpub.), thick pumice breccias appear to be absent in most of the Manus Basin. Isolated rhyolitic explosive eruptions have occurred from emergent vents with thick (up to 35 m) fall deposits and their resedimented equivalents locally preserved (Reynolds et al., 1980).

(4) Unlike the northern Central Volcanic Complex massive sulfide mineralisation in the Manus Basin is hosted in dacitic to basaltic lavas and domes and their resedimented autovolclastic facies.

7.4.2 Hokuroku Basin, Japan
The Hokuroku Basin is part of the Green Tuff Belt in northern Honshu, Japan (Fig. 5.4). The Hokuroku Basin is a 30 by 30 km submarine depositional basin containing approximately 1 km of volcanic stratigraphy (Urabe, 1987). Interpretations for the depth below sea level during deposition have varied from shallow marine (<500 m) to deep water (2000-3000 m) (Guber and Merrill, 1983; Ohmoto and Skinner, 1983).

Volcanic facies in the central Hokuroku Basin include coherent and autobreccia facies of basaltic and rhyolitic lavas, rhyolitic domes and cryptodomes, pumiceous hyaloclastite and primary pyroclastic facies (Horikoshi, 1969; Tanimura et al., 1983; Urabe, 1987). Volcanic activity is interpreted to have been initiated by andesitic effusive volcanism (Ohmoto, 1978). This was followed by the eruption of basaltic and rhyolitic lavas and associated volcanlastic facies of the Hotakizawa Formation (Horikoshi, 1969; Ohmoto, 1978; Nakajima, 1988). Overlying the Hotakizawa Formation is the Yukisawa Formation dominated by rhyolitic lavas, domes and pumice-rich facies (Ishikawa, 1983; Urabe, 1987; Yamagishi, 1987). Many of the pumice-rich facies in the Yukisawa Formation are pumiceous hyaloclastite (Chapter 5, Fig. 5.5), autobreccia and dome-related pyroclastic deposits (Horikoshi, 1969; Takahashi and Suga, 1974; Urabe, 1987). Other pumice-rich facies are thick, regionally extensive units interpreted to be associated with the development of felsic calderas (Ohmoto, 1978). This was followed by a second phase of mainly effusive bimodal activity (Kagoya Formation) interpreted to reflect post-caldera effusive volcanism (Horikoshi, 1969; Ohmoto, 1978; Ohmoto and Takahashi, 1983). The final phase includes pumice breccia and pumice-rich sandstone (Shigenai Formation) interpreted to reflect the resurgence of caldera activity (Ohmoto and Takahashi, 1983). The products of explosive eruptions are intercalated with black mudstone and rhyolitic lavas (Nakajima, 1988). Pyroclastic facies in the Green Tuff Belt have been interpreted to be emplaced directly from large submarine explosive eruptions by mass-flow processes and water-settling from suspension (Fiske and Matsuda, 1964; Urabe, 1987).

Diagenetic alteration in the Hokuroku Basin has produced a series of flat-lying alteration zones which grade vertically from glass at the top, to clays, zeolites and albite at depth (Iijima, 1978). Hydrothermal alteration and Kuroko mineralisation are associated with felsic volcanism in the Yukisawa Formation (Ohmoto and Takahashi, 1983) and formed synchronous with the regional diagenetic alteration (Iijima, 1978; Ohmoto, 1978).

Similarities between the Hokuroku Basin and the northern Central Volcanic Complex include:

(1) the size of the depositional basin (30 x 30 km) is a similar order of magnitude to the preserved size of the northern Central Volcanic Complex (30 x 7 km);
(2) both successions contain thick accumulations (up to 400 m) of felsic lavas, domes and intrusions overlain by large-volume, regionally extensive pumice-rich facies;

(3) lavas and domes with pumiceous margins and carapaces of in situ, clast-rotated and resedimented pumiceous hyaloclastite are common;

(4) shallow syn-volcanic intrusions with peperitic margins are abundant;

(5) pumice-rich breccia, sandstone and siltstone in these successions are interpreted to have been deposited from syn-eruptive water-supported mass flows and settling from suspension in the water column;

(6) in both successions, massive sulfide mineralisation is associated with the initial phase of widespread felsic volcanism and is interpreted to have formed within a submarine caldera;

(7) there is a spatial association between rhyolitic intrusions in the footwall and mineralisation (Horikoshi, 1969; Ohmoto and Skinner, 1983, Allen, 1993 unpub.);

(8) hydrothermal alteration and mineralisation were synchronous with regionally extensive diagenetic alteration.

Despite these similarities there are some important difference:

(1) unlike the northern Central Volcanic Complex, the volcanic succession in the Hokuroku Basin is truly bimodal, comprising tholeiitic basalts and calc-alkaline rhyolites (Dudas et al., 1983; Urabe, 1987);

(2) the stratigraphic thickness of the Central Volcanic Complex is interpreted to be considerably greater (1.5-3 km) than in the Hokuroku Basin (~1 km);

(3) exhalites (Kosaka Tetsusekiei beds) associated with Kuroko deposits in the Hokuroku Basin are absent in the northern Central Volcanic Complex (cf. Kalogeropoulos and Scott, 1983);

(4) diagenetic alteration zones in the Hokuroku Basin are considerably thinner (<300 m thick) than those in the northern Central Volcanic Complex (>2 km).

7.5 Implications for other comparable volcanic successions

The results of this study are relevant to volcanic and alteration facies analyses in other comparable ancient submarine volcanic successions:

(1) Research into modern seafloor volcanism has mainly been limited to andesitic and basaltic volcanic centres. As a result, knowledge of submarine felsic volcanism is largely inferred from subaerial eruptions or interpreted from ancient volcanic successions, such as the Central Volcanic Complex, which are now exposed on land. This study contributes to our understanding of volcanic facies and processes in
submarine environments, in particular, pumice-rich facies that include pyroclastic pumice breccia, pumiceous hyaloclastite and pumiceous peperite (Chapters 3 and 5). In addition, it considers the processes and constraints on eruption and emplacement of these facies.

(2) The identification of pumiceous peperite helps to constrain the age relationships and timing of pumiceous intrusions in the volcanic succession (Chapter 5). The pumiceous nature of this facies suggests that the confining pressure was insufficient to inhibit vesiculation and hence emplacement occurred at shallow levels in the sub-surface under a thin cover of sediment. Thus, pumiceous peperite formed contemporaneously or nearly contemporaneously with the host sedimentary facies. The interpretation that pumice clasts may form in the shallow sub-surface is particularly important when trying to identify seafloor positions.

(3) Studies on the diagenetic changes which occur in originally glassy felsic volcanic successions are almost exclusively limited to well sorted, fine-grained volcaniclastic facies (Hein and Scholl, 1978; Taylor and Grommé, 1981; Klein and Lee, 1984; Hay and Goldsmid, 1987; Marsaglia and Tazaki, 1992; Tazaki and Fyfe, 1992; Torres et al., 1995). In mafic volcanic successions, post depositional alteration studies are generally limited to seafloor alteration (Bonatti, 1965; Hay and Iijima, 1968; Honnorez, 1978; Zhou and Fyfe, 1989). Despite the overprinting regional metamorphism and deformation, this study has made a significant advance in unravelling the diagenetic evolution of a partly glassy and partly crystalline, rhyolitic to basaltic, submarine volcanic succession. Diagenetic textural and compositional changes in coherent and volcaniclastic facies have been assessed (Chapter 6). The diagenetic system is interpreted to have evolved through a series of recognisable stages with increasing depth of burial (Chapter 6 and 7). The most dramatic changes in texture occurred in poorly to moderately sorted, coarse volcaniclastic facies, such as pumice breccia and hyaloclastite. These changes have significant impact on the porosity, permeability and fluid pathways within the volcanic succession over time.

(4) The discrimination of hydrothermal alteration from regional alteration styles such as diagenetic alteration and metamorphism is an important exploration tool. In this study, criteria that can be applied to other volcanic successions have been developed to help discriminate between the products of hydrothermal and diagenetic alteration (Chapter 6).

7.6 Possible directions for further research

(1) Research in this thesis suggests that the contact between the Hercules Pumice Formation and the overlying White Spur Formation is a major stratigraphic boundary. This boundary reflects a change in source and it may also reflect a considerable time gap at the end of the Central Volcanic Complex. A detailed study of this contact, specifically the variation in facies along the top of the Host-rock member, the relative stratigraphic position and significance of the dacitic pumice breccia units and the conformable or disconformable nature of the contact, is necessary. This would also help to constrain the depth below sea level of eruption and deposition of the underlying Hercules Pumice Formation and the timing of mineralisation relative to emplacement of the White Spur Formation.

(2) In this thesis, it has been proposed that the thick, laterally extensive units of pumice breccia
represent the syn-eruptive products of a large explosive eruption which were deposited as caldera fill. However, the margins of this theoretical caldera have not been recognised. Further facies and structural analysis are required to help constrain the margins of the caldera and determine the size of the Central Volcanic Complex depositional basin. In particular, recognition of caldera margin faults, syn-volcanic growth faults, ring dykes, coarse syn-eruptive breccias and correlative extracaldera volcanic facies may help constrain the margins of the caldera (cf. Smith, 1960; Ohmoto and Takahashi, 1983; Walker, 1984). Pumice breccias that are exposed further north from the study area should be mapped in detail.

(3) The Sterling Valley Volcanics are interpreted to be a succession of tholeiitic basalts and andesite erupted early in the Central Volcanic Complex. This suggests that the Sterling Valley Volcanics are the lowest known stratigraphic unit of the Mount Read Volcanics and as such they have important implications for the tectonic and volcanic development of the Mount Read Volcanics. A comprehensive study of the geochemical signature of the Sterling Valley Volcanics is warranted to characterise the magmatic source and affinities.

(4) Many of the diagenetic and alteration facies described in this thesis (Chapter 6) contain similar mineral assemblages comprising sericite, chlorite, calcite and quartz. Whole-rock geochemical changes associated with the alteration facies have been assessed, however no attempt has been made to chemically differentiate similar minerals in different alteration facies. Detailed mineral chemistry of the alteration minerals may help to distinguish between the products of diagenetic and hydrothermal alteration and metamorphism. In addition, sericite, chlorite and calcite in hydrothermal alteration facies associated with mineralisation should be compared with sericite, chlorite and calcite in other hydrothermal alteration facies.

(5) Dramatic changes in porosity and permeability have been inferred to have occurred in the volcanic succession during diagenesis. These changes have significant implications for fluid pathways. Any attempt to quantify or model fluid flow in a submarine volcanic succession needs to consider changes in porosity and permeability over time. However, only limited data on the porosity and permeability of poorly and moderately sorted glassy volcaniclastic facies is available (ie. Davis, 1966, 1969; Sparks, 1978; Houghton and Wilson, 1989; Thomas, et al., 1994). A quantitative study on the porosity changes in pumice breccia during different successive stages of diagenesis in an active diagenetic system is required.

7.7 Summary

Volcanic and alteration facies analyses in the northern Central Volcanic Complex provided information about eruption and emplacement mechanisms of the volcanic facies and the post-depositional changes that have occurred. The northern Central Volcanic Complex comprises the intercalated products of two large volcanic centres, one mafic and the other felsic, deposited in a submarine environment. Tholeiitic andesitic and basaltic lavas, sills and syn-eruptive volcaniclastic facies in the Sterling Valley Volcanics are interpreted to represent a large mafic volcano. Coeval calc-alkaline dacitic activity occurred late in the formation of the Sterling Valley Volcanics. The felsic volcanic complex was initially dominated by effusive eruptions which produced a thick succession of rhyolitic and dacitic coherent and autoclastic
facies (Mount Black Formation). Onlapping the margins of this volcanic complex are sparse units of crystal-rich sandstone derived from an extrabasinal source. Dominantly effusive eruptions were followed by large explosive eruptions that produced thick ignimbrite-like pumice-rich units (Kershaw and Hercules Pumice Formations). These syn-eruptive pumice-rich units are broadly proximal, and were probably deposited within a caldera. Caldera subsidence was synchronous with or immediately after eruption. Sparse turbidites derived from a distal rhyolitic volcanic centre are intercalated with the top of the proximal pumice-rich facies. The final stage in the formation of the Central Volcanic Complex was the emplacement of rhyolitic and dacitic lavas, domes and syn-volcanic intrusions. During this stage, syn-eruptive pumice-lithic clast-rich facies derived from fault scarp collapse, resedimentation of autoclastic facies or dome-related explosions were deposited onto the caldera floor. Significant hydrothermal alteration and mineralisation are interpreted to have commenced during this final stage and continued during deposition of the overlying White Spur Formation.

The northern Central Volcanic Complex was subsequently modified by diagenesis, hydrothermal alteration, metamorphism and deformation. Diagenesis evolved through a series of recognisable stages, which advanced rapidly through the stratigraphy as overlying strata accumulated. These stages included the coating of original surfaces and infilling of primary porosity by new minerals, the dissolution and alteration of glass synchronous with compaction, and the replacement of any remaining glass and clays by more stable minerals. These stages are most clearly defined in originally porous facies, such as pumice breccia, where the original porosity is dramatically reduced during diagenesis. Precipitation of cements was synchronous with significant hydrothermal alteration and massive sulfide mineralisation at Rosebery and Hercules. Cambrian diagenetic and hydrothermal alteration assemblages and textures were overprinted by Devonian regional metamorphic mineral assemblages and tectonic foliations. Further textural and mineralogical modification resulted from the intrusion of granite at the southern end of the Rosebery ore deposit.

The volcanic facies, the palaeogeographic setting and part of the post-depositional history of the northern Volcanic Complex are similar to other younger and less deformed submarine volcanic successions in the Manus Basin and the Hokuroku Basin.